Depths in a Day - a New Era of Rapid-Response Raman-based Barometry using Fluid Inclusions

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Key Points:

* Raman-based fluid inclusion barometry can determine magma reservoir depths within one day of sample receipt.
* Our global melt inclusion volatile compilation indicates the fluid inclusion method is viable at many frequently active and hazardous mafic volcanic systems.
* The fluid inclusion barometric method is demonstrated to be viable for monitoring purposes.

Word count: Abstract 148, Text 4427 (including abstract). Figure count: 3

**Plain Language Summary**

Monitoring volcanic eruptions is crucial for nearby populations. Volcanic observatories regularly use geophysical and geochemical data to understand eruption dynamics and guide emergency response. Magma storage depth has been identified as high-priority information that is not currently obtainable via petrological methods on timescales useful to monitoring efforts. Though this has not been formally tested, it has been suggested that CO2-rich fluid-inclusions could provide this information in near real-time. We performed a rapid-response simulation for the September 10th 2023 Kīlauea eruption and compiled fluid compositions at the point of vapour saturation from global melt inclusion data to formally test the method’s potential as a monitoring tool. Using Raman spectroscopy, we obtained robust magma storage depths within one day of sample-receipt, demonstrating the value of this method for observatories in highly active and hazardous mafic volcanic systems (e.g. Iceland, Hawai’i, Galápagos, East African Rift, Réunion, Canary Islands, Azores, Cabo Verde).

Abstract

Rapid-response petrological monitoring is a major advance for volcano observatories to build and validate models of the plumbing systems that supply eruptions in near-real-time. The depth of magma storage has recently been identified as high-priority information for volcanic observatories that is not currently obtainable on timescales relevant to eruption response. To address this deficiency, we performed a rapid-response simulation for the September 2023 eruption of Kīlauea. We show that Raman-based fluid-inclusion barometry can robustly determine reservoir depths within a day of receiving samples - a transformative timescale for decision making that has not previously been achieved by petrological methods. Additionally, our global melt-inclusion compilation for which we calculated fluid composition at the point of vapour saturation demonstrates the robustness of the fluid-inclusion method at many of the world’s most active and hazardous mafic volcanic systems (e.g. Iceland, Hawai’i, Galápagos, East African Rift, Réunion, Canary Islands, Azores, Cabo Verde).

1. Introduction

Volcano observatories increasingly use data collected from erupted lava and tephra samples in near-real-time to obtain information about the magmatic plumbing system to help inform decision-making during volcanic crises (Gansecki et al., 2019; Pankhurst et al., 2022; Re et al., 2021). Most work so far has focused on the chemistry of erupted lavas and crystal cargoes (Pankhurst et al., 2022) to gain insight into changing melt composition and rheological properties (e.g., Gansecki et al., 2019). However, up until now, petrological monitoring has been unable to address the high-priority question– *Where is the magma coming from?* (Re et al., 2021). At well-monitored volcanoes, such information can be used to draw analogies to previous eruptive episodes associated with specific storage reservoirs (e.g., vigour, pathway, or length of eruption), and to help interpret geophysical signals of ongoing activity. At poorly-monitored volcanoes, where there may be no prior constraints on magma storage geometry (P. E. Wieser, Kent, Till, & Abers, 2023), depths of storage are a vital parameter to begin interpreting unrest associated with a new episode of eruptive activity (Pritchard et al., 2019). For example, the return of eruptive activity at Kīlauea in 2020 was accompanied by many questions about how the magmatic plumbing system had changed following the summit collapse in 2018 (Lynn et al., 2024).

Melt inclusion (MI) barometry, a widely popular petrological method to determine storage depths from volatile contents, takes months to years to complete (Re et al., 2021). While mineral barometry can be implemented faster than melt inclusion measurements (only requiring electron probe microanalysis (EPMA) measurements on eruptive minerals), it is imprecise (P. E. Wieser, Kent, & Till, 2023), and therefore would only be able to constrain magma storage to very broad regions of the crust (e.g., stored in the crust vs. below the Moho). Mineral thermobarometry also has poor applicability at active volcanoes such as Kīlauea or Mauna Loa where the only major silicate phase in most lavas is olivine, the chemistry of which is not pressure sensitive, and where a precision of 1–2 km is needed to distinguish between storage reservoirs (K. R. Anderson & Poland, 2016; Baker & Amelung, 2012).

Recent developments have shown that Raman-based barometry of CO2-rich fluid inclusions (FI) provides an alternative to popular petrological barometers, with much smaller uncertainties than mineral barometry, and requiring far less time and resources than MI analyses (Dayton et al., 2023; C.L. DeVitre & Wieser, 2024). This method uses spectral features of CO2 fluids to calculate a CO2 density using an instrument specific calibration (Charlotte L. DeVitre et al., 2021). Along with an estimate of entrapment temperature, this density is converted into an entrapment pressure using a CO2 Equation of State (EOS, Fig. 1). One major advantage of this method is that the conversion of CO2 density to pressure is relatively insensitive to the choice of entrapment temperature, a parameter which may not be known at the onset of a new eruptive episode (Fig. 1a-b). If we perform EOS calculations using the lower and upper limit of liquidus temperatures for olivine-saturated melts erupted at Kīlauea volcano throughout its history (~1100 and 1350 ˚C; (C.L. DeVitre & Wieser, 2024), the change in pressure is at most ~20 %, which corresponds to ~0.2 – 0.4 km at depths representative of the shallower Halemaʻumaʻu reservoir (1-2 km), and ~0.6-1 km at the depths of the deeper south Caldera reservoir (3-5 km, Fig. 1b and Fig. S3-S6). These errors are of similar magnitude to those associated with the conversion of pressures to depths through an estimate of crustal density (an issue affecting all petrological barometers).

A number of recent studies have speculated that fluid inclusion barometry can be performed quickly enough to be useful for near real-time volcano monitoring (e.g., refs (Dayton et al., 2023; Zanon et al., 2024)). However, this is the first study to rigorously assess just how quickly FI depths can be obtained from erupted material, and whether these timescales are short enough to have utility as a petrological monitoring tool. The CONVERSE Hawai‘i Scientific Advisory Committee (K. M. Cooper et al., 2023) specifically recommended that key science questions should be identified, and pre-planning science activities performed, to facilitate more rapid implementation across a broader scientific group during hazardous eruptions. Performing these simulations during relatively small, low hazard eruptions (as here) or as hypothetical simulations (e.g., Andrews et al., 2019) is vital to ensure that any bottle necks are ironed out so that we are as prepared as possible for the next large volcanic crisis (Dietterich & Neal, 2022). Importantly, this simulation revealed that rapid response fluid inclusion work in collaboration with academic institutions was not taxing on observatory staff and thus that this method can be employed during future eruptions.

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***Figure 1. Sensitivity of the fluid inclusion barometry to Temperature and*** *. (a) CO2 density vs Pressure for different magmatically relevant entrapment temperatures at Kīlauea using the EOS of* Span & Wagner, (1996)*. 1100 and 1350 ˚C are the lower and upper limit of liquidus temperatures for olivine-saturated melts erupted at Kīlauea volcano throughout its history. 1150 ˚C was the temperature used for EOS calculations during day 1 and 2 of the simulation, 1240 ˚C is the rounded mean and median of all measured temperatures in our final dataset. (b) Close-up of panel a. Grey boxes show the depth range of Kīlauea magma storage inferred from FI and MI barometry as well as geophysics* (C.L. DeVitre & Wieser, 2024; Lerner et al., 2024)*. HMM= Halema’uma’u reservoir, SC = South Caldera reservoir. Stars show hypothetical FI trapped at HMM and SC reservoirs with our initial guessed temperature (1150 ˚C) and error bars representing 1σ uncertainty from Monte Carlo simulations using a temperature uncertainty of ±125 ˚C (). (c) CO2 density vs Pressure at 1150 ˚C for various molar proportions of H2O in the exsolved fluid phase () using the mixed H2O-CO2 EOS of* (Duan & Zhang, 2006)*. Note that a small discontinuity is observed at 200 MPa due to parameter values being switched at this pressure* (Yoshimura, 2023)*. (d) Close-up of panel c. Stars show hypothetical FI trapped at HMM and SC reservoirs with our initial guessed temperature (1150 ˚C), inferred from the 2018 LERZ MI -P relationship from* C.L. DeVitre & Wieser, (2024) *and error bars representing 1σ uncertainty from Monte Carlo simulations using an uncertainty of ±0.1*

2. Timeline of Rapid Response Simulation

The eruption onset of Kīlauea volcano on September 10, 2023 provided an unprecedented opportunity to test the validity and speed of the fluid inclusion method, given that depths of the two main magma storage regions (Halemaʻumaʻu – HMM at 1-2 km and South Caldera – SC at 3-5 km) at this volcano have been well constrained by various independent geophysical and petrological methods, including prior FI barometry (C.L. DeVitre & Wieser, 2024; Lerner et al., 2024). Tephra samples representing the first ~14 hours of the September 2023 eruption were collected by Hawaiian Volcano Observatory (HVO) geologists on September 12 and mailed to UC Berkeley on September 15th ( Fig. S1). A schematic of the workflow and detailed timeline is available in the supplement (Fig. S1).

Our simulation started on September 20 at 9 am PST (Day 1), the morning after sample receipt ( Fig. S1). We used a production-line style workflow involving two undergraduates, a 1st year graduate student, a post-doc, and an assistant professor, with stations for crushing and sieving, mineral picking, FI preparation, sample cataloguing, and analysis. We crushed and sieved tephra, picked olivine crystals (size fractions 0.5-1 and 1-2 mm), and mounted them in CrystalBondTM\* to search for FI. By ~2 pm PST, we collected our first Raman spectra, and by ~7 pm PST, we had calculated CO2  densities from 16 FI using a calibration of the relationship between CO2 density and Fermi diad splitting distance (Charlotte L. DeVitre et al., 2021; C.L. DeVitre & Wieser, 2024). We calculated pressures using the pure CO2 EOS of Span & Wagner (1996). At the time of our simulation, it was challenging to perform EOS calculations considering the possible presence of H2O in the exsolved fluid due to a lack of publicly available software running on modern operating systems. However, recent work by Yoshimura (2023) identified errors in the published equations for the H2O-CO2 EOS of Duan & Zhang (2006) and provided open-source C code meaning that such calculations can be now be performed in DiadFit. Using estimates of from previously published melt inclusion data at Kīlauea (C.L. DeVitre & Wieser, 2024; P. E. Wieser et al., 2021), calculated pressures would be ~10% higher than originally reported to HVO if the CO2-H2O EOS had been used ( Fig. 2d, Fig S1 in supplementary materials). This does not affect the interpretation of our results, as the shift is far smaller than the pressure offset between the HMM and SC reservoir.

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***Figure 2. Evolution of results over 4 days.*** *a) Schematic model of Kīlauea’s plumbing system, indicating reservoir depths determined by geophysics and prior petrological work (HMM- Halemaʻumaʻu; SC – South Caldera). b) Day 1 FI data, as reported to HVO. FI-derived depths are consistent with the estimated depths of the HMM reservoir. Kolmogorov-Smirnoff tests indicate that September 2023 FI are recording significantly shallower depths than FI (critical D = 0.22, stat = 0.24, pval=0.016) and MI (critical D = 0.22, stat = 0.41, pval=3.51e-06) from the 2018 lower East Rift Zone (LERZ) eruption, which required a contribution from the SC reservoir* (C.L. DeVitre & Wieser, 2024; P. E. Wieser et al., 2021)*. 1LERZ 2018 melt inclusion data* (P. E. Wieser et al., 2021)*; 2LERZ 2018 fluid inclusion data* (C.L. DeVitre & Wieser, 2024)*. c) Day 2 data, as reported, confirmed a likely dominant role for the Halemaʻumaʻu reservoir. A conservative degassing filter was applied (SO2/CO2 peak ratio < 0.1). d) Day 4 data, as reported. Means were taken for repeated analyses of single FI and more stringent data filters, FI-specific temperatures, and a more appropriate crustal density model (~2300 kg/m3 with a normal error distribution of 100 kg/m3) were applied. We also filtered using an SO2/CO2 peak ratio < 0.22. Error bars correspond to uncertainties propagated using Monte Carlo simulations and olivine Fo equilibrium field is calculated based on Glass EPMA data collected on September 11, 2023 (see Supplementary Information S1 Appendix). The shifted histogram ‘H2O effect’ shows the effect of H2O corrections on pressures recalculated using inferred from MI data* (C.L. DeVitre & Wieser, 2024; P. E. Wieser et al., 2021)*.*

For the first and second days, we assumed an entrapment temperature of 1150 ˚C for all FI, based on geothermometric estimates of previously erupted liquids (C.L. DeVitre & Wieser, 2024; Gansecki et al., 2019). On day 4, we calculate entrapment temperatures for each FI using the host forsterite content measured by Energy Dispersive Spectroscopy (EDS) (C.L. DeVitre & Wieser, 2024), yielding temperatures spanning 1182–1307 ˚C. The average error induced by our initial assumption of 1150 ˚C is only ~7% (with a maximum offset of only 12 %). While crystallization temperatures at Kīlauea are relatively well constrained relative to other volcanic systems, using similar regression methods to those employed by C.L. DeVitre & Wieser (2024) relating liquid compositions to host olivine contents, it should always be possible to constrain temperatures within ~100 K at different volcanic systems using host mineral chemistry.

On days 1 and 2, pressures were converted into depths using the crustal density model of (Ryan, 1987) parameterized by Lerner et al. (2021). We shared the resulting histogram ( Fig. 2a-b) of storage depths with HVO collaborators showing that crystals, and thus magma, were likely coming from the shallower Halemaʻumaʻu reservoir of Kīlauea (HMM on Fig. 2a-b). It worthwhile to note that the number of FI reported on Day 1 (N=16) is comparable to many melt inclusion studies, which often aim for ~20 MI per sample but frequently report fewer. For example, (Lerner et al., 2021) reported only 9 MI from the 2018 eruption with sufficient data to calculate saturation pressures (counting MI with glass major element contents and H2O contents, and glass CO2 measurements if there was no bubble, and glass + bubble measurements if a bubble was present). Using the same criteria, (Aster et al., 2016) only reported 13 MI complete measurements from Lassen Peak.

We also had an additional ~20 FI fully prepared and catalogued for analysis by the end of Day 1. On Day 2, these 20 FI were analysed, while additional FI were prepared and catalogued. After analysis of ~15 crystals hosting FI, these crystals removed from CrystalBondTM\* and placed on tape to make an epoxy mount. Epoxy was poured at the end of Day 2. By ~8:30 pm PST on Day 2, we shared an updated histogram of 46 FI pressures and depths from 28 crystals, confirming the dominant contribution of the Halemaʻumaʻu reservoir ( Fig. 2a and c). On Day 3, while waiting for the epoxy to fully set, we finished analysing the remaining prepared FI. Then we polished the mount and began cataloguing the regions of crystals which were closest to each FI on which to perform EDS analyses. On Day 4, olivine forsterite contents (Fo = 100\*Mg/(Mg+Fe) molar) were determined by EDS, providing a framework to further interpret the plumbing system ( Fig. 2d). The Fo content of an olivine is a function of MgO and FeO in the liquid and the Ol-Liq partitioning coefficient (KD). Thus, the Fo contents of the host olivine close to each FI can be used to assess the calculated storage depth in its broader petrographic context (e.g., distinguishing high-Fo olivines which crystallize from more primitive melts from low Fo olivines forming in more evolved melts). This olivine forsterite content can also be used to estimate the likely entrapment temperature of each fluid inclusion (see (C.L. DeVitre & Wieser, 2024)) for performing EOS calculations, rather than having to use a uniform temperature as on Day 1-2. We recalculated all FI pressures on Day 4 using fluid inclusion specific entrapment temperatures from our EDS data.

Our results on Day 4 clearly show that the majority of FI were entrapped at ~1–2 km below the surface ( Fig. 2d), which aligns well with the depths of the Halemaʻumaʻu reservoir interpreted from geophysics (K. R. Anderson et al., 2019; K. R. Anderson & Poland, 2016; Baker & Amelung, 2012), MI barometry (Lerner et al., 2021; P. E. Wieser et al., 2021), and FI barometry (C.L. DeVitre & Wieser, 2024; Lerner et al., 2024). While the greater number of analyses from data processed on Day 2 and Day 4 certainly enhance the story, it is notable that depths calculated on Day 1 fall within final proposed storage reservoir depths. Rapid EDS analyses of olivine Fo contents close to each FI reveal that olivine crystals grew from a wide range of melt compositions. It is interesting to note that FI in the cores of high-Fo (e.g., >86) olivine crystals return pressures indicative of the shallower Halemaʻumaʻu reservoir, given that it has been suggested based on previous eruptions that these high-Fo olivine crystals predominantly grow in the deeper South Caldera reservoir (SC on Fig. 2a) where high MgO melts are thought to reside (Rosalind T. Helz et al., 2014; Lerner et al., 2024; Pietruszka et al., 2015, 2018; P. E. Wieser et al., 2019). We suggest three possible scenarios to explain the relatively shallow pressures documented in high-Fo olivine crystals:

1) FI in high-Fo olivine crystals were entrapped within the South Caldera reservoir and then transported into the Halemaʻumaʻu reservoir, where the FI re-equilibrated to lower pressures prior to eruption over shorter timescales than would be required to reset the host Fo content.

2) High-MgO melts were injected into the Halemaʻumaʻu reservoir, where high-Fo olivine crystallized and trapped FI at shallow depths (Lerner et al., 2024).

3) Complex skeletal growth of olivine crystals during extensive undercooling (Welsch et al., 2013) could mean that high-Fo olivine cores which initially grew in the SC reservoir texturally evolved and trapped lower pressure FI in the Halemaʻumaʻu reservoir.

We think that scenario 1 is unlikely given the that FI from the 2018 lower East Rift Zone eruption appear not to have re-equilibrated despite stalling in the Halemaʻumaʻu reservoir for up to 2 years (C.L. DeVitre & Wieser, 2024; Mourey et al., 2023), and our models of FI re-equilibration indicate <10% change in pressure over ~ 2 yrs. Current data does not allow us to resolve scenario 2 vs 3, but this eruption could provide an opportunity to explore this further, such as through detailed Phosphorous mapping in olivine around FI (as performed for melt inclusions by Esposito *et al.*, 2023). Regardless of the exact mechanism, our FI pressures indicate that erupted crystal cargo experienced storage at Halemaʻumaʻu reservoir depths prior to eruption, and thus this was the most probable reservoir supplying magma to the surface in the Sept 2023 eruption.

3. Broader applicability of the method

The use of a pure CO2 EOS results in an underestimate of the entrapment pressure of fluid inclusions if there was H2O in the fluid at the time of inclusion entrapment (Fig. 2). At Kīlauea, melt inclusion data indicates that the exsolved fluid phase is ~90% CO2 at pressures corresponding to the HMM reservoir, and >95% CO2 at pressures indicative of the SC reservoir (C.L. DeVitre & Wieser, 2024; P. E. Wieser et al., 2021). As discussed above, the effect of is small on calculated pressures presented here ( Fig. 2d). However, to assess the utility of the fluid inclusion method for rapid-response petrology globally, it is necessary to evaluate contents, and the effect on calculated pressures from fluid inclusions.

We compiled published melt inclusion data from all over the world, spanning many different tectonic settings (Fig. 3). We calculate using the solubility model MagmaSat (Ghiorso & Gualda, 2015), implemented in VESIcal (Iacovino et al., 2021). We show the distribution, median, 25th and 75th quantiles of calculated for 4069 melt inclusions with SiO2 < 57 wt%, MgO < 16 wt% and Saturation Pressure > 20 MPa on Fig 4b-c, coloured by tectonic setting. For each volcano, there is a clear correlation between and pressure with values taking an uptick to very high values at shallow pressures (Supporting Fig. S9a-i) as melt inclusions are trapped during enhanced degassing of H2O upon ascent. Thus, in this compilation, the median and 25th quantiles are likely most representative of in the main magma storage region. We stress the importance of considering when determining the suitability of this method to a particular system given that the pressure correction tends to be more significant at higher entrapment pressures. For example, if we consider an of 0.1 (the median of our FI dataset at Kīlauea and a commonly assumed in deep storage systems), the pressure correction goes from <15% at pressures < 220 MPa (~10 km) to ~20% at 700 MPa (~30 km). Naturally, the correction is even more significant if is greater than 0.1 (e.g., for =0.2, the correction is 25-30% at P<150 MPa and ~50% at P = 700 MPa; Fig 1). We plot on Fig 4a the median for magmas with SiO2 < 57 wt% and MgO < 16 wt% around the world.

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*Figure 3.* ***Global compilation of in the exsolved fluid phase from melt inclusion data for Continental Rift, Continental Intraplate, Alkaline and Tholeiitic Ocean Island Basalt (OIB, see Fig S7) Mid-Ocean Ridge and Subduction Zone volcanoes (details and references in the supplement). Data is filtered to SiO2 < 57 wt%, MgO < 16 wt% and Saturation Pressure >20 MPa (supplement for details).*** *(a) World map coloured by Median of the melt inclusion suites, circles indicate Glass-only MI data and stars those for which CO2 has been constrained by Raman. (b) Boxplot of for melt inclusion suites plotted on panel a. Boxplots show the median, Q1 (25th quartile), Q3 (75th quartile) and whiskers mark the last datapoint before Q3+1.5\* (Q3-Q1) and the first datapoint after Q1-1.5\*(Q3-Q1). Violin plots show the density distribution of all the data and are coloured according to tectonic setting. (c) Boxplot of showing only MI suites which constrained Total CO2 by Raman spectroscopy. EAR – East African Rift, GSC - Galápagos Spreading Center, NAR – North Atlantic Ridge, JdFR – Juan de Fuca Ridge, GR – Gakkel Ridge, MAR – Mid-Atlantic Ridge, EPR – East Pacific Rise, IBM – Izu Bonin Mariana.*

It is also noteworthy that the vast majority of melt inclusion suites in this compilation did not measure CO2 in the vapour bubble, meaning that the total CO2 content of the inclusion has been underestimated, and overestimated. This can be demonstrated by comparing values at volcanoes where there some studies with Raman measurements and some without (Supporting Fig. S9c EAR, Fig. S9h Kamchatka, and Cascades). Thus, Fig. 3c shows a compilation only using melt inclusions where bubble CO2 was accounted for by Raman spectroscopy.

Both compilations demonstrate that subduction zones record much higher globally than Mid-Ocean Ridge basalts, ocean island basalts, continental rifts and intraplate volcanoes. It is also interesting that within hotspot and intraplate settings, regions with tholeiitic compositions (e.g. Iceland, Hawai’i, Galápagos, Réunion, Deccan Traps) generally have lower values than regions with more alkaline magmas (e.g., Canary Islands, Azores, Cabo Verde, Supporting Fig. S9). This likely represents the lower melt extents in alkaline settings, and the possibility of more volatile-rich sources (e.g.,DeVitre *et al.* 2023).

Thus, while rapid response fluid inclusion barometry is highly applicable to active volcanic regions such as Hawai’i, Iceland, East African Rift (EAR), Galápagos, Réunion, Cabo Verde, and the Canary Islands, it is not appropriate in subduction zones such as Alaska, Kamchatka, or Central America. Interestingly, although there are only two studies with Raman data in the Cascade arc (Aster et al., 2016; Venugopal et al., 2020a), it is noteworthy that the highest pressure inclusions have values <0.2. This may indicate that in dryer subduction zones such as the Cascades, fluid inclusions may have some utility for the most mafic, CO2-rich magmas.

To increase the accuracy of rapid-response petrological monitoring during future eruptions, it should be a priority to perform melt inclusion studies accounting for vapour bubble CO2 in more volcanic systems worldwide, given the large offsets between studies accounting bubbles and those which do not in for space (Fig. 3). Ideally, for maximum accuracy of fluid inclusions as a rapid response method, it should be a priority to determine approximate trends in -pressure space for a given volcanic system or region during times of quescience. This will allow assessment of the suitability of the fluid inclusion method and permit appropriate corrections for the complexities of mixed fluids without requiring melt inclusion work during each eruptive episode. In systems with no prior constraints, our observations of correlations between alkalinity and can provide a first order assessment of appropriate values to use. We note that once arc magmas are excluded from the compilation, even if is entirely unconstrained, fluid inclusion barometry is still more accurate than other methods such as mineral-melt thermobarometry.

4. Conclusion

This simulation shows that magma storage depths can be determined within a day of receiving samples, with modest resources and personnel requirements (e.g. no overnight shift work, with normal semester teaching and class schedules). For example, sample preparation was carried out using transmitted-reflected light microscopes from the University of California teaching collection, only using a research-grade microscope for sample cataloguing. Raman spectrometers are widely available at many universities, given that it is a popular technique in many other fields, such as material sciences, physics, chemistry, and biology, and the W-filament SEM used for EDS analyses to get olivine Forsterite contents has been around for 15 years (See Supplementary Information S1 Appendix). Importantly, this simulation shows that rapid response work in collaboration with universities was not taxing on observatory staff, particularly considering the usefulness of information provided. This means this methodology can be employed during future eruptions to help observatories deduce the geometry of the plumbing system supplying magma, adding a crucial information for interpreting activity (Re et al., 2021), without retracting from other essential duties during eruption responses. For example, during the 2018 Kilauea eruption, HVO’s near real time chemical monitoring with bulk rock ED-XRF identified the appearance and disappearance of many magma batches (Gansecki et al., 2019), fluid inclusion barometry could have linked these distinct chemical signatures to different storage regions, addressing the questions of scientists and residents alike. Similarly, the return of eruptive activity at Kīlauea in 2020 after this event was accompanied by many questions about how the magmatic plumbing system had changed following the summit collapse in 2018 (Lynn et al., 2024). FI barometry would have been a critical addition to understanding the eruption and the system.

Our global compilation of values shows that Raman-based FI barometry has utility as a rapid-response petrological monitoring method at many of the world’s most active and hazardous basaltic volcanoes (e.g., Galápagos, Réunion, Azores, Canary Islands, Iceland, Cabo Verde). As our understanding of exsolved fluid compositions improves as more studies account for CO2 held within vapour bubbles, it is likely that the applicability of this method may expand to even more volcanic systems (e.g. to drier arc magmas such as those found in the Cascades).

Overall, fluid inclusion barometry is a broadly applicable, and adds valuable quantitative storage depth information that provides a key advancement for volcano observatories that utilize near-real-time geochemical monitoring to better understand eruptions as they unfold (See overview–(Re et al., 2021); Hawai‘i–(Gansecki et al., 2019); La Palma–(Pankhurst et al., 2022); Fuego–(Liu et al., 2020); Italy–(Corsaro & Miraglia, 2022)).

5. Acknowledgements

Any use of trade, product, or firm names is for descriptive purposes only and does not imply endorsement by the U.S. Government.

**5.1. Funding**

National Science Foundation grant EAR 2217371 (PW and CLD)

Berkeley Rose Hills Innovator Program (PW, CLD, AB, BR and RR)

**5.2. Author contributions**

Author contributions for lab work are shown on Fig. S1. CLD and PW wrote the paper. CLD, PW, AR, BR, and AB prepared tephra, picked olivine, found FIs, catalogued them, mounted them, and conducted Raman analyses. CLD compiled the melt inclusion dataset. CLD and PW performed all spectral fitting, data processing, and figure making, with schematic cartoons shown in Fig. S1 from AB. JG developed the Mg/Fe calibration for the EDS detector and MG performed EDS analyses with help from JG. KJL, DTD, NID and KMM collected samples, processed them in Hilo, provided eruption context and edited the manuscript. KJL and DD prepared the glass mount and did the EMPA glass analyses.

**5.3. Competing interests**

Authors declare that they have no competing interests.

6. Data availability

All data are made available in the Supplementary Information associated with the publication. We include detailed materials and methods (S1 Appendix), complete processed fluid inclusion dataset (S2 Dataset), the global melt inclusions dataset (S3 Dataset), a compilation of microphotographs of the FI and crystals that were used for navigation only during the simulation (S3 FI Image Compilation) and a record of emails reporting results to HVO and tracking receipts related to sample shipment (S4 Email and tracking record). All raw data and Jupyter notebooks are also stored on Github at the following link: <https://github.com/cljdevitre/2023_Kilauea-rapid-response-simulation>. The Github repository will be archived on Zenodo upon acceptance.

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*Geophysical Research Letters*

Supporting Information for

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S4 FI Navigation Image Compilation

S5 Email and tracking record

All other raw data (spectra, metadata, FI images) as well raw Jupyter processing notebooks can be found on a Github repository (<https://github.com/cljdevitre/2023_Kilauea-rapid-response-simulation>) which will be archived at Zenodo upon acceptance.

**Introduction**

The supporting information contains the materials listed above.

**Text S1. Sample collection and preparation**

Tephra samples (USGS code KS33-588) representing the first ~14 hours of the September 10, 2023, eruption of Kīlauea volcano were collected by Hawaiian Volcano Observatory (HVO) geologists on September 12 and shipped on Friday September 15 at ~5 pm HST (Fig. S1). This tephra was erupted from a fissure which opened at 15:36 local time on September 10 (~22 minutes after the eruption started, between 15:13 and 15:14 local time) and ceased erupting between 06:16 and 06:18 am local time on September 11. Following receipt of the samples at the University of California, Berkeley on Tuesday, September 19, material was processed in a jaw crusher in the VIBE lab which was thoroughly cleaned the week before and the morning of the simulation, and then sieved into >2 mm, 1–2 mm, and 0.5–1 mm size fraction. Crystals were picked from the 1–2 mm and 0.5–1 mm size fraction using three different binocular microscopes (one of which had the ability to cross polars). Then, crystals were individually mounted in CrystalBond™ on glass slides and progressively polished with 1200-2500-10000 grade wet and dry paper to find FI. Grains containing FI were then passed onto a team member on a research grade scope to take reflected and transmitted light images to aid with Raman navigation. These images were pasted into a Google slides document so all lab personnel at UC could access them immediately (images are compiled in supplement S5 FI Image Compilation as an example).

A lava flow sample (USGS code KS33-587) was collected in a molten state and quenched with water at 6:30 AM HST on the 11 of September 2023. The sample was entirely glassy, and fragments were mounted in a 1” epoxy round and polished for microprobe analysis.

A diagram of a machine

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***Figure S1. Workflow of the study, all times on this figure are Pacific Standard Time (PST).*** *Stick people show the contribution of individual team members, to indicate the total time associated with each step. We note that AB was trained in these procedures during the simulation and all reports were sent to HVO prior to 5:30 pm Hawaii Standard Time (HST), thus allowing for decision making for the following day.*

Text S2. Raman analyses

Raman spectra were acquired using a WiTec Alpha 300R Raman spectrometer at the Department of Earth and Planetary Science at the University of California, Berkeley. The relationship between CO2 density and spectral features was determined from a gas calibration cell following the methods of(C.L. DeVitre & Wieser, 2024). All spectra were acquired from samples heated to 37 ℃. Spectra were processed and corrected for drift using the Python tool DiadFit v0.0.73. We report ratios of SO2 to CO2 peak areas. We filtered the final dataset for SO2/CO2 ratio < 0.22 (Fig. 1d), to ensure use of the pure CO2 EOS was valid (Given an instrumental efficiency of 1 for our Raman, this corresponds to <10 mol% SO2). No fluid inclusions contained carbonate peaks nor any gaseous species other than CO2. We calculated densities from the Raman-measured separation of the Fermi diad using the appropriate calibrated density equations for our instrument in DiadFit (P. E. Wieser & DeVitre, 2023). For the final dataset (Day 4), we took a mean of duplicate analyses, and calculated pressures using the EOS of (Span & Wagner, 1996) using an entrapment temperature estimated from the Fo content close to the FI (C.L. DeVitre & Wieser, 2024). Entrapment depths in Fig. 1d were calculated using a constant crustal density of 2300 kg/m3 and a normally distributed 1**σ** error of 100 kg/m3. Error in the CO2 density for each FI was determined from the error in each peak fit, the Ne line drift correction model, and the densimeter (P. E. Wieser & DeVitre, 2023). We used a 40℃ error for temperature (C.L. DeVitre & Wieser, 2024). We propagated these three sources of uncertainty in FI depths using MonteCarlo simulations implemented in DiadFit v 66. In total we analyzed 62 FI hosted in 31 olivine crystals. Our workflow is detailed in Fig. 1 of the main text. Pictures of each FI and host crystal are available in the repository linked at the beginning. We note here that the initial data reported for Days 1 and 2 did not account for repeated analyses (1 repeated FI in Day 1 and 6 in Day 2; we took a mean of repetitions on Day 4), pressures were calculated using an estimated entrapment temperature of 1150˚C (C.L. DeVitre & Wieser, 2024; P. E. Wieser et al., 2021), and depth was calculated using the crustal density model in (Lerner et al., 2021; Ryan, 1987).

**Text S3. Epoxy mount making and polishing**

After Raman analysis, crystals were removed from CrystalBondTM\* using a hotplate and placed in Acetone. They were then mounted on double-sided sticky tape with their polished side down. EpoFixTM\* resin was used to impregnate the samples in a Cast-N-Vac vacuum pourer. After curing, the epoxy mount was polished using an EcoMet30 automatic polisher, with 9, 3, and 1 um diamond pastes. A reflected light map and image of each crystal was taken using the Raman microscope to aid SEM sample navigation. The location within each FI in the reflected light image was cataloged so the Scanning Electron Microscope (SEM) operator knew where to analyze to obtain an approximate Fo content for each FI.

**Text S4. EDS analysis**

Samples were carbon coated to an approximate thickness of 25–30 µm for EDS analysis. Chemical data for each host crystal in the proximity of each FI was determined using a Zeiss EVO MA-10 SEM and a single AMETEK EDAX 10 mm2 detector at the University of California, Berkeley. The beam was rastered over a 30-by-30 µm area for ~75–80s (a live time of 60 seconds with ~30% dead time). For all analyses we used an accelerating voltage of 20 kV and a 30 µm aperture, giving an approximate beam current of 5 nA. EDS data reduction was performed using an in-built standardless quantification routine (including a ZAF matrix correction), alongside pre-determined “Standardless Element Coefficients” (SECs). The SECs act as correction factors for each element in the standardless quantification routine and have been determined through several years of repeat analyses of multiple different silicate standard materials and glasses. This method returns an estimate for the relative abundance of each element in the analyzed material and, if a normalization to 100% is assumed, can be used to return semi-quantitative chemical analysis of elemental or oxide weight percent values. However, for the purposes of this study we simply focused on the relative abundance of Mg and Fe in the EDS analyses to calculate the Fo content of the olivine host crystals. Furthermore, by calculating the molar Si/(Mg+Fe) ratio of each analysis we were able to provide a stoichiometric check of data quality: we obtained an average Si/(Mg+Fe) ratio of 0.497±0.006 on Kīlauea olivine crystals, close to the ideal value of ~0.5. Precision and accuracy were determined through repeat measurements of the San Carlos and Springwater olivines, which were not used as part of the standard quantification routine. We estimate the precision and accuracy of the method using repeat analyses of secondary standards (5 at start, 5 at end of day), which have Fo contents similar to our samples (see supplementary dataset S3). The Smithsonian-preferred Fo content (Jarosewich et al., 1980) for the San Carlos secondary standard is 90.2 Fo, and we obtained a mean of 89.84±0.07 Fo units. For Springwater, the preferred value is 82.4 Fo, and we obtained a mean of 82.1±0.2 Fo. We also analyzed a Kīlauea olivine crystal previously measured on the USGS Menlo Park EPMA. The average Fo content obtained at Menlo Park was 87.8±0.1 Fo units, and at the University of California, Berkeley SEM, 88.5±0.1. It should be noted that such offsets also exist between different EPMA labs (P. E. Wieser, Kent, Till, Donovan, et al., 2023). Considering these probable differences, we compared data acquired at Stanford University to that obtained at Cambridge University on the olivine crystals of (P. E. Wieser et al., 2021). The difference observed amounts to ~0.62 units at ~82 Fo and 0.78 units at 90 Fo (C.L. DeVitre & Wieser, 2024). Thus, EDS errors are within uncertainty of offsets between different microprobe labs.

Text S5. EPMA analysis of glasses

Major and minor element analysis of glass from USGS sample KS33-587 was done using the U.S. Geological Survey’s JEOL 8530F microprobe at the California Volcano Observatory. The samples were run over midnight between September 23 and 24th, 2023, in 1hr 37 minutes (9/23/2023 23:06 - 9/24/2023 0:43). A total of 20 total analyses were conducted, corresponding to 4 analyses per grain in 5 grains. These samples were run as part of a pre-booked session for other samples that started on September 20th. We note however that CalVO microprobe personnel later informed us that it would be possible in the future to request immediate access for eruption response if needed. This would mean that a glass mount could be prepared in 24 hrs after sample collection, shipped within 2 days from HVO to CalVO and analyzed on the probe on day 4 post field collection (calibration takes 2-3 hrs and analysis ~2 hours).  We also note that if these measurements did not exist at the time of the simulation, we could have used the EDS-SEM method to get the Mg# of the host glass to calculate the equilibrium olivine Fo content shown on Fig. 1 (The only reason we need this EPMA data). EDS measurements on the matrix glass were within 1-2 Mg# units of EPMA measurements – far smaller than the uncertainty associated with calculating an equilibrium olivine content based on uncertainty regarding the choice of olivine-liquid KD model at Kīlauea.

Microprobe glass analyses used 15 kV accelerating voltage and a 10 µm beam with a 10 nA current. Peak counting times were 45 s for S, Cl, and F, 40 s for Ti, P, and Mn, 20 s for Si, Ca, Fe, Al, and Mg, and 10 s Na and K (backgrounds were measured on both sides of the peak for half the peak counting times). Standards were VG2 basaltic glass (USNM 111240/52;ref (Jarosewich et al., 1980)) for Si, Mg, and Al, Kakanui Pyrope Garnet (USNM 143968) for Fe, and Al, Wollastonite for Ca, Tiburon Albite for Na, MnO3 for Mn, TiO2 for Ti, Orthoclase OR-1A for K, Wilburforce Apatite (USGS-M105731) for P, Barite for S, Sodalite for Cl, and MgF2 for F. Two-sigma relative precision, based on two analyses of VG-2 glass (before and after lava sample was run), are 0.19 wt% for SiO2, 0.15 wt% for Al2O3, 0.003 wt% for TiO2, 0.27 wt% for FeO, 0.009 wt% for MnO, 0.006 wt% for MgO, 0.04 wt% for CaO, 0.11 wt% for Na2O, 0.02 for K2O, 0.04 for P2O5, 0.07 for SO3, 0.0001 for Cl, and 0.002 for F. X-ray intensities were converted to concentrations using standard ZAF corrections(Armstrong, 1988). Analyses with totals <99.0 wt% or >100.5 wt% were rejected. Reported analyses are an average of four replicate points on individual glass fragments.

Text S6. Manuscript Writing Timeline

The study presented here was formulated into a letter over days 4 and 5 (September 23-24th), sent to our co-authors on Day 6 (September 25th) and submitted for review to Nature Geoscience on Day 8 - September 27th, 2023 (see S6\_Email\_and\_tracking\_record for email confirmations), one week after we begun the simulation. Unfortunately, despite our prompt submission, we did not receive a rejection notification until a month later, owing to editorial delays. The rejection, based on the grounds of 'lack of appeal for the broader Geoscience community', was surprising, given the significant interest of the Geoscience community in hazard mitigation. We proceeded to submit the manuscript to PNAS on October 31st, 2023 (see S6\_Email\_and\_tracking\_record) who rejected the manuscript on November 15th (see S6) on similar grounds with the editor comments as follows: “This is indeed an interesting real-time procedure but may be too specialized for PNAS”. After submitting to Journal of Petrology on November 16th, 2023, we received a rejection from JPET on January 19th, 2024, based on concerns from reviewers of temperature sensitivity of EOS and the lack of applicability to subduction zones (S6). We have addressed these concerns with the new Fig. S1 and Fig. 3 and further details in the supplement and body of the article and believe the new manuscript has significant global applicability, hence our submission to Science Advances. Regardless it is evident that the delays in the editorial and publishing process constitute possibly the biggest bottleneck for reporting findings to the community, particularly given USGS collaborations mean that results cannot be posted as preprints prior to peer review.

Text S7. Identifying and Resolving Bottlenecks

The yellow stars on Fig. S1 identify current bottlenecks in the process that could be easily improved.

***Star 1 – Shipping and receiving samples***

Distributing samples to the University of California, Berkeley was not a top priority for HVO because this simulation was being attempted for the first time, and as a result, there was no guarantee of obtaining magma storage depths in a timely manner. Samples were shipped from Hilo on a Friday at ~5 pm HST. HVO was asked to ship samples to a private residence under the assumption that they might arrive over the weekend. However, no packages leave Hilo after 4pm on Friday over the weekend, so the samples started their transit to California on Monday. Had the package been taken to the courier’s office on Friday morning, it would likely have arrived on Sunday. The tracking information indicated arrival on Wednesday, which is when we planned to start the simulation. However, the samples arrived at the private residence on Tuesday morning during working hours, without notification that the package had been delivered (and no one was home).

We have demonstrated that this technique adds valuable quantitative depth information that expands on HVO’s routine near-real-time chemical monitoring with bulk rock ED-XRF(Gansecki et al., 2019). Under ideal circumstances, HVO geologists would sample tephra or molten lava from the eruption on Day 1 (morning) and dry the samples in the lab on Day 1 (afternoon), dropping the samples for shipment on the evening of Day 1, which would go out early on Day 2 (as long as the drop off did not occur Friday afternoon or over the weekend). Same-day shipping from Hawaii to California is not realistic, but samples shipping Monday through Thursday mornings would allow for arrival on Day 3. Additionally, it would be possible to get samples to the University of California, Berkeley within 24 hours if someone in Hawai’i were to take a flight to San Francisco or Oakland airport with the samples, or within ~30 hours if someone based in the University of California, Berkeley flew to Hawai’i, picked up the samples, and returned home immediately.

We note that this bottleneck can be avoided entirely if observatories rely on local research expertise (for example, in collaboration with local academic research groups) and/or establish in-house workflows for this type of work. In such a case, next-day information could readily be obtained. As this is not an option for HVO, the Hawaii-California connection will serve as the fastest way to conduct this rapid-response barometry.

***Star 2 – Sample cataloging***

The WITEC Raman microscope used in this study does not have a condenser in its optical path, which can make it difficult to navigate and find FIs directly, particularly in volcanic crystals that are commonly coated in glass. The first 7–10 crystals were analyzed immediately after preparation with no navigation photos, so finding the FI on the Raman scope added some time. After AB had finished crushing, sieving, and picking, he began taking photos on a research-grade scope to help the Raman operator find the FIs they were supposed to be analyzing. Late on Day 2, when students were not available, Wieser began photographing crystals holding her phone to the eyepiece of the teaching-collection reflected light and transmitted light microscopes. This provided enough textural context to easily find FIs on the Raman (See Image Compilation S5 in the repository linked in the beginning). The main advantage of using smartphones is that the person who found each FI could identify it, rather than passing it off to another person who then must work out where the FI is before photographing it. This would greatly reduce the number of people needed for the simulation, as we almost always had one person taking photos.

***Star 3 – Epoxy impregnation***

We used EpoFixTM epoxy in our laboratory because it gives low backgrounds during SIMS analysis. After pouring the epoxy at ~7 pm, it was removed from its mount at ~9 am the next morning. The epoxy was still noticeably soft (to the extent it cracked coming out of the mold). This meant that we could not start polishing immediately. Instead, we had to wait a further ~5 hours for it to cure sufficiently to polish. If fast-curing epoxies were available, it is very possible that a team member could have stayed, and polished and cataloged the sample overnight, allowing SEM analysis on Friday (Day 3) rather than Saturday.

Text S8. Effect of H2O on calculated pressures

The exsolved fluid phase in shallow magmatic systems like Kīlauea is not pure CO2, but rather contains a proportion of H2O. Fluid inclusion studies typically assume that H2O has been lost and therefore the measured CO2 density must be corrected based on the molar fraction of H2O and molar ratios (see (Hansteen & Klügel, 2008)). With this, pressures can be calculated using a mixed H2O–CO2 equation of state. Although it was not possible to implement these calculations during our simulation, a recent paper (Yoshimura, 2023) made it possible to implement these corrections in DiadFit. We recalculated pressures for our fluid inclusions using mol fractions of H2O in the exsolved fluid calculated based on the polynomial equations for Kīlauea in ref (C.L. DeVitre & Wieser, 2024). We iterated measurements 5 times and show calculated on Fig S2; the data for iteration 5 is reported in the full FI dataset (Dataset S3) and the jupyter notebook is available in the repository linked in the beginning. We note that for all 3 days, the mean and median correction factor is ~10%. Most FI have correction factors <20%. These correction factors do not shift our FI from one reservoir to another, therefore they do not affect the interpretation of our results.

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**Fig. S2**. Cumulative probability plot of the ratio of pressures from the mixed fluid EOS and pure CO2 EOS depending on the  function applied from (C.L. DeVitre & Wieser, 2024) for each Day of the simulation.

Text. S9. Sensitivity of the EOS method for FI to entrapment temperature – Extended

We test the sensitivity of 3 equations of state for CO2 available in DiadFit: two are for pure CO2 (Span & Wagner, 1996; Sterner & Pitzer, 1994) and one for CO2-H2O mixtures (Duan & Zhang, 2006). For this, we calculated pressures using all three EOS at different magmatically relevant entrapment temperatures (for the mixed H2O-CO2 EOS we used  = 0.1) which encompass the entire range of measured ol-saturated liquid temperatures at Kīlauea volcano (1100,1150, 1240 and 1350 ˚C). Overall, neither of the three equations of state are significantly sensitive to temperature at these magmatic temperatures (Fig. S3-S6). The shaded box in Fig. S3 shows storage conditions relevant to Kilauea. In the worst case, at the depth of the HMM reservoir (1 km), the absolute uncertainty in depth due to temperature is ~0.15 km and at depths corresponding to the SC reservoir (~4 km) it is ~0.7 km. We plot in detail on Figs S4-S6 the temperature sensitivity for each of the EOS at general conditions and conditions specifically relevant to Kīlauea volcano.

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Fig. S3. % Difference in pressure (a) and absolute difference in depths (b) for pressures calculated at 1350˚C and 1150˚C using two pure CO2 EOS (SW96 (Span & Wagner, 1996), SP94 (Sterner & Pitzer, 1994)) and one mixed H2O-CO2 EOS (DZ06 (Duan & Zhang, 2006)). Shaded area a indicates the range of pressures and depth relevant to Kīlauea pre-eruptive magma storage.

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Fig. S4. Temperature sensitivity of the mixed CO2-H2O equation of state at 0.1 mol% H2O (Duan & Zhang, 2006).(a) Pressure vs CO2 density calculated for various entrapment temperatures relevant at Kīlauea volcano. 1150 ˚C was our initial fixed temperature for days 1 and 2, 1240 ˚C is the rounded mean and median of all measured temperatures in our final dataset. (b) Depth and Pressure differences induced by uncertainty in temperature. Blue curves show the % difference in pressure (or depth) and maroon curves show the absolute difference in depth in km. (c) Closeup of panel a, representing relevant PT conditions for Kīlauea volcano. Grey boxes show the depth range of the magma storage reservoirs – HMM for Halema’uma’u and SC for South Caldera – inferred from FI and MI barometry as well as geophysics (C.L. DeVitre & Wieser, 2024; Lerner et al., 2024). Stars show hypothetical FI trapped at HM and SC reservoirs with error bars representing 1σ uncertainty from Monte Carlo simulations using a temperature uncertainty of ±150 K. (d) Closeup of panel b, showing depth and pressure differences induced by uncertainty in temperature for PT conditions relevant to Kīlauea volcano. Blue curves show the % difference in pressure (or depth) and maroon curves show the absolute difference in depth in km.

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Fig. S5. Temperature sensitivity of Span and Wagner 1996 pure CO2 equation of state (Span & Wagner, 1996).

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Fig S6. Temperature sensitivity of Sterner and Pitzer 1994 pure CO2 equation of state (Sterner & Pitzer, 1994).

Text S10. Global Melt inclusion compilation

We compiled melt inclusion datasets with published major element data, H2O and CO2 concentrations in the glass from EarthChem (https://www.earthchem.org/), Georoc (https://georoc.eu/), (D. J. Rasmussen et al., 2022)), (Charlotte L. DeVitre et al., 2023) and (P. E. Wieser, Kent, Till, & Abers, 2023)) as well as other MI datasets from the literature (Chelsea M. Allison et al., 2021; Chelsea Maria Allison, 2018; A. T. Anderson & Brown, 1993; Aster et al., 2016; Auer et al., 2009; Bali et al., 2018; Benjamin et al., 2007; Bennett et al., 2019; Berlo et al., 2012; Bouvet de Maisonneuve et al., 2012; Brounce et al., 2014; Cassidy et al., 2015; Cervantes & Wallace, 2003; Colman et al., 2015; L. Cooper et al., 2022; L. B. Cooper et al., 2010; Charlotte L. DeVitre et al., 2023; Donovan et al., 2017; Druitt et al., 2016; R. Esposito et al., 2011; Famin et al., 2009; Field et al., 2012; Gennaro et al., 2019; Gleeson et al., 2022; Harris & Anderson Jr, 1983; Hartley et al., 2014; E. Hauri et al., 2021; E. H. Hauri et al., 2017; Head et al., 2011; Helo et al., 2011; Hernandez Nava et al., 2021; Hudgins et al., 2015; Iddon & Edmonds, 2020; Johnson et al., 2009; Kelley et al., 2010; Kelley & Cottrell, 2009; Koleszar et al., 2009; Lerner et al., 2021; Lloyd et al., 2013; Longpré et al., 2017; Manzini et al., 2019; Métrich et al., 2014; Miller et al., 2019; de Moor et al., 2013; L. R. Moore et al., 2021; Lowell R. Moore et al., 2015, 2018; Mormone et al., 2011; Myers et al., 2014; Plechov et al., 2015; D. J. Rasmussen & Plank, 2021; Daniel J. Rasmussen et al., 2017; Ribeiro et al., 2015; Roberge et al., 2009; Robidoux et al., 2017, 2018; Roggensack et al., 1997; Roggensack, 2001b, 2001a; Rose-Koga et al., 2012; Ruscitto et al., 2010, 2011; Ruth et al., 2016; Saal et al., 2002; Sadofsky et al., 2008; Schipper et al., 2011; Shaw et al., 2010; Sides et al., 2014; Taracsák et al., 2019; Venugopal et al., 2016, 2020b, 2020a; Vigouroux et al., 2008; Wade et al., 2006; Wallace et al., 2015; Walowski et al., 2016, 2019; Wanless et al., 2014, 2015; Wanless & Shaw, 2012; P. E. Wieser et al., 2021; Wong et al., 2023; Wysoczanski et al., 2012; Zimmer, 2009; Zimmer et al., 2010).

We calculated saturation pressures and the fraction of H2O in the exsolved fluid phase () using the MagmaSat solubility model (Ghiorso & Gualda, 2015) implemented in VESIcal (Iacovino et al., 2021). Whenever measurements were available, we used total CO2 contents calculated from mass balance for Raman measured MI bubbles, otherwise we use glass-only CO2. Also, when possible, we use post-entrapment crystallization corrected MI concentrations. As a first approximation, temperatures were calculated from major element data using the CaO liquid-only thermometer of (Rosalind Tuthill Helz & Thornber, 1987) implemented in Thermobar (P. Wieser et al., 2022) as it is not pressure sensitive and from there we calculated saturation pressures using MagmaSat. After this initial step, we recalculated temperatures using those pressures with the more appropriate ol-liq equation 22 of (Putirka, 2008), coupled to a theoretical calculation of the DMg from (Beattie, 1993), to calculate equilibrium olivine compositions so that equation 22 of (Putirka, 2008) can be used as a liquid-only thermometer, as we did not have access to olivine pair chemistry (this is function “T\_Put2008\_eq22\_BeattDMg” in Thermobar (P. Wieser et al., 2022)). This equation has recently been identified as the most robust liquid-only thermometer across a wide range of compositions (P. E. Wieser, Gleeson, Matthews, et al., 2023). After this, we recalculated saturation pressures using MagmaSat and our refined temperatures. We did not recalculate saturation pressures, temperatures or  for MI in the Cascades compilation of (P. E. Wieser, Kent, Till, & Abers, 2023). It is a known issue that MagmaSat may fail to converge and/or crash during saturation pressure calculations, which can be problematic when calculating saturation pressures on large datasets such as our >4000 MI compilation. To circumvent crashing issues, we implemented a python multiprocessing routine along with a try-except block using the python package “multiprocessing”. The dataset was first divided in small chunks of ~20 samples and assigned to each processor core (in the case of this study a Dell Inspiron 15 with i7 8 core processor) limited to a maximum of 8 processes run simultaneously (as the author’s laptop has 8 cores, so one process per core at a time). Within each ~20 sample chunk, the MI were run one by one within a try-except block such that whenever an error occurred, NaN values were returned along with the error message from VESIcal (Iacovino et al., 2021). Any MI for which MagmaSat did not converge were filtered out in the final dataset and the notebook is included in the data repository linked in the beginning. We provide both the complete unfiltered compilation, and the compilation filtered for MagmaSat errors, SiO2 < 57 wt% and MgO < 16 wt%. As discussed in previous sections, our method is not applicable to systems and magmas where  is very high, like for example in high silica magmas where H2O is concentrated due to fractional crystallization and CO2 is lost due to extensive degassing. Therefore, we plot only mafic MI with SiO2 < 57 wt% (Fig. S7), likely representative of recharge magmas regardless of tectonic setting and for which the method could be applicable to determine locations in the world where the method could be relevant. Interestingly, it is evident that  tends to be higher at alkaline intraplate volcanoes than their tholeiitic counterparts at similar pressures (Fig. S8).

We then plot  against pressure at each unique location, grouped by tectonic settings (Fig. S9a-e). It is notable that there is a marked increase in  at very low pressures (<50 MPa). We filter the dataset presented on Fig 4, and Fig S10 considering a pressure cutoff of 20 MPa, where  is very high regardless of setting.

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Figure S7. Total Alkalis vs Silica diagram for all the MI <57 wt% SiO2 and MgO <16 wt% in our compilation. Larger black edged symbols indicate the medians of specific locations. (a) Intraplate (Oceanic and Continental) volcanoes in the compilation. We separate OIB-Tholeiitic and OIB-Alkaline locations based on the medians on this plot. (b) Subduction zone volcanoes. (c) Mid-Ocean Ridge (d) Continental Rifts.

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Figure S8. TAS diagram (a) and Pressure vs  for Tholeiitic and Alkaline Ocean Island volcanoes and one continental intraplate volcano.

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Figure S9a. Pressure vs  at Tholeiitic OIB locations.

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Figure S9b. Pressure vs  at Alkaline OIB locations and sunset crater (an alkaline continental intraplate volcano).

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Figure S9c. Pressure vs  at Continental rift locations.

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Figure S9d. Pressure vs  at Mid-Ocean Ridge locations (Part 1). JdFR – Juan de Fuca Ridge, GSC – Galápagos Spreading Center, EPR – East Pacific Rise.

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Figure S9e. Pressure vs  at Mid-Ocean Ridge locations (Part 2). NAR – North Atlantic Ridge, MAR – Mid-Atlantic Ridge, GR – Gakkel Ridge.

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Figure S9f. Pressure vs  at Subduction Zones (Part 1).

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Figure S9g. Pressure vs  at Subduction Zones (Part 2).

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Figure S9h. Pressure vs  at Subduction Zones (Part 3).

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Figure S9i. Pressure vs at Subduction Zones (Part 4).

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Figure S10. Boxplots and violin plots of H2O and CO2 contents of melt inclusions in the compilation. Filtered to SiO2 < 57 wt%, MgO < 16 wt% and Saturation Pressure > 20MPa.

Data S2. (separate file)

Fluid inclusion dataset for this study.

Data S3. (separate file)

Global melt inclusion compilation (described in previous sections).

**Image Compilation S4 (separate file)**

Compilation of images used for navigation during the simulation. These are presented as they were used during the simulation as an example of what is necessary for the technique.

**Email and tracking record S5 (separate file)**

Compilation of tracking labels, email exchange with HVO personnel and editorial process.

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