**Supporting Information for**

‘Depths in a day - A new era of rapid-response Raman-based barometry using fluid inclusions’

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**This PDF file includes:**

Detailed Materials and Methods

Sample collection and preparation

Raman analyses

Epoxy mount making and polishing.

EDS analysis

EPMA analysis of glasses

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Identifying and Resolving Bottlenecks

**Other supporting materials for this manuscript include the following:**

S2 Dataset (Excel Table attached to the submission)

S3 FI Image Compilation

S4 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.

# Detailed Materials and Methods

**Sample collection and preparation**

Tephra samples (USGS code KS23-588) 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. 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 fractions 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 S3 FI Image Compilation).

A lava flow sample (USGS code KS23-587) 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.

**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 DeVitre and 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. 2d), 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. We calculated densities from the Raman-measured separation of the Fermi diad using the appropriate calibrated density equations for our instrument in DiadFit(Wieser and DeVitre, 2023). For the final dataset (Day 4), we took a mean of duplicate analyses, and calculated pressures using the EOS of Span and Wagner, (1996) using an entrapment temperature estimated from the Fo content close to the FI (DeVitre and Wieser, 2024). Entrapment depths in Fig. 2d 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 (Wieser and DeVitre, 2023). We used a 40℃ error for temperature (DeVitre and 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. Pictures of each FI and host crystal are available in the repository linked at the beginning (Image Compilation). 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 (Wieser *et al.*, 2021; DeVitre and Wieser, 2024), and depth was calculated using the crustal density model in Lerner *et al.*, (2021).

**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.

**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 S2). 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 (Wieser *et al.*, 2023a). Considering these probable differences, we compared data acquired at Stanford University to that obtained at Cambridge University on the olivine crystals of Wieser *et al.*, (2021). The difference observed amounts to ~0.62 units at ~82 Fo and 0.78 units at 90 Fo (DeVitre and Wieser, 2024). Thus, EDS errors are within uncertainty of offsets between different microprobe labs.

**EPMA analysis of glasses**

Major and minor element analysis of glass from USGS sample KS23-587 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. 2 (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.

**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 S4\_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 S4\_Email\_and\_tracking\_record) who rejected the manuscript on November 15th (see S4) 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 JPET on November 16th 2023 date, we received a rejection from JPET on January 19th 2024 date, based on concerns from reviewers of temperature sensitivity of EOS and the lack of applicability to subduction zones. Consequently, 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 means results cannot be posted as preprints prior to peer review.

**Identifying and Resolving Bottlenecks**

The yellow stars on Fig. 1 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 Fransisco 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 very hard to navigate and find FIs, 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 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.

**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 and 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 et al., 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 (DeVitre and Wieser, 2024). We iterated measurements 5 times and show *X*H2Ocalculated on Fig S1. 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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**Figure S-1** Cumulative probability plot of the ratio of pressures from the mixed fluid EOS and pure CO2 EOS depending on the *XH*2O function applied from (DeVitre and Wieser, 2024) for each Day of the simulation.

**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 (Sterner and Pitzer, 1994; Span and Wagner, 1996) and one for CO2-H2O mixtures (Duan and 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 XH2O = 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. S2-S5). At storage conditions relevant to Kīlauea volcano (shaded box in Fig S2), the absolute uncertainty in depth induced by temperature uncertainty of 100 K is less than 2 km (~15% in pressure). In the worst case, at the depth of the HM 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 S3-S5 the temperature sensitivity for each of the EOS at general conditions and conditions specifically relevant to Kīlauea volcano.

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Fig S2. % 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 and Wagner, 1996), SP94 (Sterner and Pitzer, 1994)) and one mixed H2O-CO2 EOS (DZ06 (Duan and Zhang, 2006)). Shaded area a indicates the range of pressures and depth relevant to Kīlauea pre-eruptive magma storage.

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Fig S3. Temperature sensitivity of the mixed CO2-H2O equation of state at 0.1 mol% H2O (Duan and Zhang, 2006). **Figure 1. Temperature sensitivity of the FI barometer.** (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 – HM for Halema’uma’u and SC for South Caldera – inferred from FI and MI barometry as well as geophysics (DeVitre and Wieser, 2024; Lerner *et al.*, 2024). Stars show hypothetical FI trapped at HM and SC reservoirs with error bars representing 1σ uncertainty from MonteCarlo 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 S4. Temperature sensitivity of Span and Wagner 1996 pure CO2 equation of state (Span and Wagner, 1996).

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Fig S5. Temperature sensitivity of Sterner and Pitzer 1994 pure CO2 equation of state (Sterner and Pitzer, 1994).

**Global Melt inclusion compilation**

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

We calculated saturation pressures and the fraction of H2O in the exsolved fluid phase (XH2O) using the MagmaSat solubility model (Ghiorso and Gualda, 2015) implemented in VESIcal (Iacovino *et al.*, 2021). Whenever possible, 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. Temperatures were calculated from major element data using the CaO liquid-only thermometer of Helz and Thornber (1987) implemented in Thermobar (Wieser *et al.*, 2022) given the wide range of compositions of melt inclusions in the compilation. While we recognize this may not be the most appropriate nor accurate thermometer for every scenario, MagmaSat is fairly insensitive to entrapment temperature (Ghiorso and Gualda, 2015; Iacovino *et al.*, 2021). Temperatures for Rasmussen *et al.*, (2017) were not recalculated but rather we used the author’s reported entrapment temperatures for the MI. Note that 99% of calculated saturation pressures with temperature from CaO thermometer versus those reported in the study differ by less than 4% (90% of the whole dataset < 2% different). The XH2O predicted using those two temperatures is <3.5 % different for the whole dataset. Therefore, we consider that the CaO liquid-only temperature provides a sufficiently reasonable estimate for the purpose of this study. Similarly, we did not recalculate saturation pressures, temperatures or XH2O for MI in the Cascades compilation of Wieser *et al.*, (2023b). Any MI for which MagmaSat did not converge were filtered out. We provide both the complete XH2O compilation, and the compilation filtered for SiO2 < 57 wt%. As discussed in previous sections, our method is not applicable to systems and magmas where XH2O 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 XH2O tends to be higher at alkaline intraplate volcanoes than their tholeiitic counterparts at similar pressures (Fig. S8).

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Figure S7. Total Alkalis vs Silica diagram for all the MI <57 wt% SiO2 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 XH2O for Tholeiitic and Alkaline Intraplate volcanoes.

We then plot XH2O against pressure at each unique location, grouped by tectonic settings (Fig. S9a-e). It is notable that there is a marked increase in XH2O 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 XH2O is very high regardless of setting.

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Figure S9a. Pressure vs XH2O at Tholeiitic intraplate locations.

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Figure S9b. Pressure vs XH2O at Alkaline intraplate locations.

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Figure S9c. Pressure vs XH2O at Continental rift locations.

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Figure S9d. Pressure vs XH2O 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 XH2O 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 XH2O at Subduction Zones (Part 1).

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Figure S9g. Pressure vs XH2O at Subduction Zones (Part 2).

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Figure S9h. Pressure vs XH2O at Subduction Zones (Part 3).

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Figure S9i. Pressure vs XH2O 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% and Saturation Pressure > 20MPa.

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