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Spatially heterogeneous argon-isotope systematics and apparent $^{40}\text{Ar}/^{39}\text{Ar}$ ages in perlitised obsidian

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**A B S T R A C T**

In situ laser ablation Ar-isotope analyses of variably hydrated and devitrified obsidian from the ~27 Ma Cochetopa Dome, San Juan, USA, reveal complex interplay between degassing of initial Ar and absorption of atmospheric Ar. These processes have locally modified the Ar-isotope composition of the obsidian and led to spurious, spatially-heterogeneous Ar-isotope and $^{40}\text{Ar}/^{39}\text{Ar}$ age data. Small perlite beads exhibit older apparent Ar-ages at the rims than the cores. This is interpreted as an apparent excess of $^{40}\text{Ar}$ at the rims, produced either by a) diffusion of excess $^{40}\text{Ar}$ into the bead during flushing of the lava with excess $^{40}\text{Ar}$-bearing volcanic gas, or by b) isotopic fractionation during degassing of initial Ar, causing preferential loss of $^{40}\text{Ar}$ over $^{36}\text{Ar}$ at the bead rims. The second interpretation is favoured by a relative enrichment of $^{36}\text{Ar}$ in the core of a perlite bead along a microlite-free (poorly degassed) flow band, and by a lack of age variation in a larger, fresh, well-degassed perlite bead. These isotopic gradients were later overprinted during glass hydration by absorption of Ar with near-atmospheric composition, resulting in elevated $^{36}\text{Ar}$ and reduced radiogenic $^{40}\text{Ar}$ yields at the rims of perlite beads.

These complex interactions essentially represent the mixing of three distinct Ar reservoirs: initial trapped Ar that may or may not be fractionated, an isotopically atmospheric Ar component introduced during hydration, and radiogenic $^{40}\text{Ar}$. Such reservoir mixing is the underlying reason for poor correlations on Ar-isotope correlation diagrams and the difficulties in validating the composition of the non-radiogenic Ar component. We thus suggest that high $^{36}\text{Ar}$ yields are a combination of the incomplete degassing of initial (possibly magmatic) Ar and the gain of Ar during interaction between the obsidian and meteoric/atmospheric fluids. Our analyses emphasise the challenging nature of $^{40}\text{Ar}/^{39}\text{Ar}$ dating obsidian samples, but also point to possible solutions by careful sample characterisation and selection of highly degassed samples.

**1. Introduction**

Relative and absolute chronological data is of utmost importance in interpreting the products of volcanic eruptions and how they affected the local, regional and wider environment. As isotope-dating techniques improve, so does the precision on radiometric ages, allowing precise correlations between volcanic eruptions and climate change events to be extended to prehistoric eruptions and those not represented in ice cores (c.f. Clausen et al., 1997; Gao et al., 2008; Sigl et al., 2015). Such correlations are particularly important because they facilitate a deeper understanding of both the potential environmental impact of volcanic eruptions, and the timescales of past climate change, as well as postulated feedbacks between volcanism and climate (e.g. Huybers and Langmuir, 2009; Rampino and Self, 1992). Precise dating of large volcanic eruptions that can be recognised in Arctic and/or Antarctic ice cores such as the Younger Toba Tuff (Mark et al., 2014; Storey et al., 2012) allows calibration of Quaternary climate records. However, the products of smaller volcanic eruptions can be useful in interpreting localised environmental conditions, especially at volcanic systems with temporally-variable snow and ice cover, such as in Chile (Lachowycz et al., 2015; Mee et al., 2009, 2006), Iceland (McGarvie et al., 2007; Stevenson et al., 2009, 2006; Tuffen et al., 2010) and Antarctica (Smelle et al., 2014, 2011a, 2011b). Precise dates for this type of volcanic deposit could provide a wealth of data on the timing and timescales of local environmental changes, such as ice sheet growth and retreat, which reflect broader climatic events (McGarvie, 2009). A number of recent studies highlight the potential benefits of this combination of age and volcanic/environmental facies data (Flude...
followed by a stress corrosion process, in which addition of H2O in
addition of K2O (Flude et al., 2010, 2008; Denton et al., 2012; von Aulock et al., 2013), but can continue at
temperatures below 400 °C devitrification to rather lower temperatures (300 °C; Befus, 2016). Initial
processes affecting obsidian during and after cooling. Obsidian is a volcanic glass that forms by rapid cooling of silica rich
melt. It may form by chilling of lava in contact with air, water or ice, as highly welded facies in pyroclastic deposits, and as chilled margins of shallow silicic intrusions. Devitrification of obsidian can occur over a wide
temperature range. Above and around Tg, at temperatures of 800–600 °C, spherulite and globulite growth converts obsidian to
anhydrous masses of sanidine, cristobalite and Fe-Ti oxides (Castro et al., 2008; Gardner et al., 2012; Watkins et al., 2008), but most Icelandic magmas are of tholeiitic or transitional-
alkaline lineage (Jónasson, 1994), making K-bearing phenocrysts notably scarce.

Obsidian, silicic volcanic glass, is however, relatively common, often containing several percent K2O (Flude et al., 2010, 2008; Jónasson, 1994; McGarvie, 1984; McGarvie et al., 2007; Owen et al., 2012; Stevenson et al., 2006), but most Icelandic magmas are of tholeiitic or transitional-
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Obsidian, silicic volcanic glass, is however, relatively common, often containing several percent K2O (Flude et al., 2010, 2008; Jónasson, 1994; McGarvie, 1984; McGarvie et al., 2007; Owen et al., 2012), and most recent attempts at dating Icelandic volcanics have employed 40Ar/39Ar dating of obsidian, albeit with varying degrees of success. Other recent studies have shown that 40Ar/39Ar dating of obsidian could be applied to a wide range of localities and research problems (Lipman and McIntosh, 2008; Morgan et al., 2009). However, first standardisation and optimisation of 40Ar/39Ar dating techniques as applied to obsidian is necessary, requiring understanding of the processes affecting obsidian during and after cooling.

Perlitisation involves the ingress of external water into obsidian via the formation of curving fractures, which intersect to create beads of non-hydrated glass surrounded by zones of strong H2O enrichment along concentric fractures (e.g. Denton et al., 2012, 2009; Friedman et al., 1966; Jezek and Noble, 1978; von Aulock et al., 2013). Initial fracturing is attributed to cooling contraction (Denton et al., 2012), and followed by a stress corrosion process, in which addition of H2O to crack tips triggers a volume change in the glass that drives further crack advance. Perlitisation is thought to occur at elevated temperatures ~ between the glass transition (Tg ~ 600–700 °C) to ~ 300–400 °C (Denton et al., 2012; von Aulock et al., 2013), but can continue at ambient temperature, albeit with drastically lower rates of diffusive hydration. Perlitisation can induce ion exchange, with local enrichment in K2O and depletion in Na2O corresponding with zones of hydration along perlite fractures (Jezek and Noble, 1978).

Significant difficulties in achieving accurate and precise K-Ar and 40Ar/39Ar ages on obsidian have been encountered (Flude et al., 2010; Kaneoka, 1972; Morgan et al., 2009; Vogel et al., 2006). In this paper we summarise the current understanding of the difficulties in dating obsidian using the 40Ar/39Ar method, and begin to investigate ways in which the technique might be improved through more complete understanding of 40Ar/39Ar systematics in obsidian, including refined sample characterisation and selection. We have used variably hydrated and devitrified obsidian from the San Juan volcanic field, Colorado, USA, and performed in-situ Ar-isotope analyses and infra-red spectroscopic analyses to measure the distribution of Ar-isotopes and water concentrations in the various obsidian microtextures.

1.1. 40Ar/39Ar dating of obsidian – the problems

K-Ar and 40Ar/39Ar dating often rely on the assumption that Ar-isotopes within a sample are the product of mixing between two end-member reservoirs: radiogenic 40Ar*, derived from radioactive decay of 40K, and a single trapped component, representing the argon content of the sample at the time it cooled through the closure temperature for argon. A basic age determination assumes that the trapped component is atmospheric (40Ar/39Ar = 298.5 - Lee et al., 2006) and uses the abundance of 36Ar to correct the measured 40Ar value for trapped air. The Ar-isotope composition of the trapped component can be characterised by carrying out multiple measurements on the same sample and plotting the results on an isotope correlation diagram (isochron). Many studies of 40Ar/39Ar dating on a range of materials find that the trapped component is indeed atmospheric (McDougall and Harrison, 1999). Trapped Ar isotopes with 40Ar/39Ar > 298.5 are, however, common, indicating the presence of excess 40Ar in the sample (Kelley, 2002). In these cases, plotting of an isochron can determine the true 40Ar/39Ar age and characterise the isotopic composition of the trapped argon, but only when the system represents mixing between radiogenic argon and a single reservoir or trapped, excess 40Ar-bearing argon.

Observed difficulties in K-Ar and 40Ar/39Ar dating obsidian can be generally classified as one of four problems:

1. Post-cooling hydration may mobilise alkali metals and release Ar, resulting in loss of 40Ar* and/or K (Cerling et al., 1985; Kaneoka, 1972, 1969). Water-derived H+ substitutes for Na+ or K+, facilitating charge balance and allowing the alkalis to migrate out of the glass structure (Cerling et al., 1985), while alteration of the glass structure promotes loss of argon. Loss of 40Ar* results in spuriously young apparent Ar-ages, while loss of K results in spuriously old apparent Ar-ages.

2. Many studies have noted high proportions of 36Ar, presumed to be a trapped atmospheric component. This large, trapped atmospheric argon component dilutes the 40Ar* signal and reduces age precision. Although atmospheric Ar entrapment occurs in other volcanic minerals, including micas, these are rarely noted in the literature and it appears particularly problematic for obsidians (Flude et al., 2010; Morgan et al., 2009; Vogel et al., 2006). High atmospheric Ar abundances may result from incomplete degassing of the melt during eruption (assuming an initial trapped argon component that is isotopically atmospheric), the incorporation of meteoric fluid into the obsidian during eruption, cooling and/or hydration, and via microcrack generation during sampling and sample preparation (Ballentine and Barfod, 2000). Distinguishing these trapped argon sources has thus far received little scientific attention and models for identifying these different components do not currently exist.

3. Non-radiogenic and non-atmospheric 40Ar (excess 40Ar) may be present in obsidian samples (Flude et al., 2010), a common problem for all fields of 40Ar/39Ar dating (Kelley, 2002). Failure to identify and account for excess 40Ar leads to spuriously old apparent ages. It is thought to originate from equilibration in an environment with a greater than atmospheric 40Ar/39Ar ratio, and so may be particularly problematic for primitive melts with a large proportion of mantle volatiles, and for volcanic systems set in old continental crust where 40Ar* from old, K-rich host rocks becomes assimilated into magma as excess 40Ar. Excess 40Ar is identified by plotting multiple analyses (either stepped heating or single grain fusions) on an isotope
correlation diagram (isochron) where the intercept on the y-axis represents the composition of the non-radiogenic, i.e. trapped, Ar. In simple systems the intercept will have an atmospheric value \((^{36}\text{Ar}/^{36}\text{Ar} = 295.5)\) if there is no excess \(^{40}\text{Ar}\), or, if excess \(^{40}\text{Ar}\) is present, \(^{40}\text{Ar}/^{36}\text{Ar}\) will be > 295.5. However, samples containing excess \(^{40}\text{Ar}\) may produce isochrons with an atmospheric intercept if either the excess \(^{40}\text{Ar}\) is unevenly distributed through the sample (Sherlock and Arnaud, 1999), or if the sample also contains a large proportion of atmospheric Ar – i.e. three-component mixing between radiogenic \(^{40}\text{Ar}^{36}\text{Ar}\), trapped gas with excess \(^{40}\text{Ar}\) and atmospheric Ar (Kuiper, 2002) (Fig. 1). In addition, the spread in \(^{40}\text{Ar}/^{36}\text{Ar}\) values from obsidian stepped heating data may be insufficient to produce an isochron (Morgan et al., 2009), preventing the fingerprinting of the non-radiogenic Ar component (Fig. 1). Some studies have identified obsidian samples containing trapped argon with \(^{40}\text{Ar}/^{36}\text{Ar} < 295.5\) – i.e. they contain an apparent excess of \(^{36}\text{Ar}\) (Morgan et al., 2009; Vogel et al., 2006). If not accounted for, this will lead to spuriously young apparent ages due to over-correction of atmospheric \(^{40}\text{Ar}\). Morgan et al. (2009) attributed low \(^{36}\text{Ar}/^{36}\text{Ar}\) ratios to kinetic mass fractionation of Ar-isotopes during incomplete equilibration of the obsidian with the atmosphere during quenching. Similarly, Brown et al. (2009) suggested a two-stage isotope fractionation mechanism, with atmospheric argon initially being fractionated via Soret diffusion within a gaseous environment experiencing a strong temperature gradient (e.g. air in a crack in an obsidian lava flow), followed by further kinetic fractionation as the argon diffuses into the obsidian.

No empirical data for \(^{36}\text{Ar}\) diffusivities are available to compare with that of \(^{40}\text{Ar}\). The abovementioned studies have used Graham’s law and the kinetic theory of gases to estimate the maximum difference in diffusivities between Ar-isotopes, which provides an upper estimate of the ratio of diffusivities \((0.949 \text{ for } ^{36}\text{Ar}/^{36}\text{Ar})\):

\[
\text{Diffusivity } A = \sqrt{\frac{\text{Mass of } A}{\text{Mass of } B}}
\]

\[
\text{Diffusivity } B = \sqrt{\frac{\text{Mass of } B}{\text{Mass of } A}}
\]

They noted that the co-variation of \(^{40}\text{Ar}/^{36}\text{Ar}\) with \(^{38}\text{Ar}/^{36}\text{Ar}\) is broadly consistent with this level of diffusive fractionation (Morgan et al., 2009). However, the kinetic theory of gases holds true for single ideal gases only; it is an approximation for diffusion of gases through air but unlikely to be fully representative of the diffusion of a gas through a silicate network. Furthermore, recent studies of the diffusion of noble gases (Amalberti et al., 2016) and other elements through silicate melts, glasses and other mineral lattice networks have identified diffusive

identification of the fractionation processes leading to excess \(^{36}\text{Ar}\) in obsidians and their effect on \(^{40}\text{Ar}/^{36}\text{Ar}\) systematics is revealing crucial insights, but many problematic issues persist. Firstly, experimentation using a MAP 215-50 mass spectrometer (Morgan et al., 2009; Vogel et al., 2006) cannot resolve true \(^{36}\text{Ar}\) from interfering contaminants \(^{37}\text{Cl}\) and \(^{12}\text{C}\). In these studies \(^{37}\text{Cl}\) measurements were not reported and it was not discussed whether measured \(^{36}\text{Ar}\) abundances were representative of true \(^{36}\text{Ar}\) concentrations; contaminants at mass 36 will be a particular problem for samples with low \(^{36}\text{Ar}\) contents and for geologically young samples (as most obsidians are). Secondly, it has been proposed that Ar will diffuse into cooling obsidian because the melt Ar content is lower than that of the atmosphere, but this is problematic as the atmospheric partial pressure of Ar is extremely low at \(\sim 9.3 \text{ mbar}\). Ar solubility data from diffusion experiments on rhyolitic melts at high Ar partial pressures (e.g. Carroll and Stolper, 1993) indicate likely Ar insolubility at atmospheric conditions. If true, rhyolitic melt would diffusively degas any remaining Ar, rather than resorb it from the atmosphere.

Furthermore, even if extrapolation of Ar solubility from laboratory to natural conditions were inappropriate, and Ar solubility at atmospheric pressure were non-negligible, the volume of rhyolitic melt/obsidian affected by atmospheric Ar resorption would be strongly limited by the diffusion rate of Ar in silicate melts and glasses. In order for significant diffusive resorption to occur, an unrealistic path length may be required (for equivalent arguments for H\(_2\)O diffusive degassing from fragmented rhyolitic melt see Castro et al., 2012). Fig. 2 illustrates Ar diffusion profiles calculated for diffusion of Ar into obsidian at temperatures from 850 °C (a typical extrusion temperature) to 550 °C. The diffusion profiles were calculated using Fick’s second law, and assuming an Arrhenian relationship with an activation energy \((E)\) of 144 kJ mol\(^{-1}\) and a pre-exponential factor \((D_0)\) of \(-6.16 \text{ m}^2\) to calculate diffusivity \((D)\) at different temperatures (Behrens, 2010a, 2010b; Carroll and Stolper, 1991; Carroll, 1991). Even if negligible cooling occurred over ten days, Ar diffusion would only affect the outer \(\sim 1 \text{ mm}\) of lava surfaces (the upper carapace, plus fractures and bubbles that are permeably connected, Castro et al., 2012).

Constraints on the cooling rate of natural obsidian within lavas or pyroclastic material are predominantly gained via relaxation geospectrometry, where the thermal behaviour of the sample around the glass transition temperature \((\sim 600–700 °C)\) is used to reconstruct its initial rate of cooling (e.g. Gottsmann and Dingwell, 2001). Published results
display considerable variability, both between and within individual lava bodies, ranging from 1 to over 100 K per day (Gottsmann and Dingwell, 2002, 2001; Seaman, 2013; Wilding et al., 1995). This was demonstrated for an obsidian lava on Lipari, Italy where the upper and lower margins of a flow ramp cooled at > 140 °C per day, while the inner core cooled at only ~59 °C per day (Gottsmann and Dingwell, 2001). There is thus generally a coincidence between the portions of a lava body that cooled the most rapidly and those most likely to have interacted with the atmosphere, due to greater proximity to cooling surfaces. Thus, portions of obsidian most likely to interact with air will be those that have shorter lengthscales for diffusive Ar uptake. Plausible cooling timescales within the temperature range over which significant Ar diffusion occurs (erupted at 750–900 °C and cooling to ~500 °C) are therefore in the range 2–500 days. It is therefore unlikely that temperatures in lava flows would be maintained sufficiently long to produce diffusion profiles exceeding a few hundred microns in length. This is especially true for the model of Brown et al. (2009) where the presence of a crack formed in the brittle regime requires the obsidian to have already cooled below the glass transition temperature of ~600–700 °C e.g. (Seaman, 2013). In reality, ephemeral fracture systems may develop in obsidian above Tg (Tu = 6.16 m² s⁻¹ and assuming a spherical geometry (Behrens, 2010b; Carroll and Stolper, 1991; Carroll, 1991). It takes 10 days at eruption temperatures for Ar to diffuse the same distance as water in ~2 h.

![Image](https://example.com/fig2.png)

**Fig. 2.** Calculated concentration profiles for Fickian diffusive gain of water (Castro et al., 2005) and Ar in obsidian. Ar diffusion parameters: $E = 144$ kJ mol⁻¹, $\log D_0 = -6.16$ m² s⁻¹ and assuming a spherical geometry (Behrens, 2010b; Carroll and Stolper, 1991; Carroll, 1991). It takes 10 days at eruption temperatures for Ar to diffuse the same distance as water in ~2 h.

Consequently, only a small proportion of the total volume of obsidian is likely to have diffusively exchanged Ar with the atmosphere, which challenges models of atmospheric Ar contamination of obsidian. Ar $^{40}$Ar/$^{39}$Ar analysis is typically conducted on chips of fresh, non-hydrated obsidian far from exposed surfaces, connected bubbles and pervasive fractures, there is generally little scope for significant post-eruptive Ar diffusion into measured samples.

Evidently, large gaps in our understanding of the $^{40}$Ar/$^{39}$Ar systematics in obsidian remain, which need to be addressed before robust age determination is possible. Many of these problems are exacerbated by the fact that obsidians are usually geologically young, being metastable supercooled fluids that seldom persist beyond a few millions of years before converting to devitrified crystalline assemblages. The low abundance of radiogenic $^{40}$Ar* in such young samples results in greater uncertainties in the $^{40}$Ar/$^{39}$Ar system than in older samples.

2. Sample and methodology

Obsidian from Cochetopa Dome (San Juan, USA) is some of the oldest known on Earth (~27 Ma; (Lipman and McIntosh, 2008)) and forms fresh-looking nodules (perlitic beads/“apache tears”), surrounded by a hydrated matrix, and extensively devitrified, flow-banded material. Cochetopa dome samples thus afford an excellent opportunity to make detailed measurements on fresh, hydrated and devitrified obsidian, in material sufficiently old to reasonably resolve radiogenic $^{40}$Ar* from trapped $^{40}$Ar. Sample OOL-31A was previously analysed by Lipman and McIntosh (2008) by laser stepped heating of fresh obsidian from the cores of large, fresh perlite beads and gave a plateau age of 26.9 ± 0.1 Ma (1 standard deviation), using the decay constants and isotope abundances of Steiger and Jäger (1977) and a Fish Canyon Sanidine fluence monitor age of 28.02 Ma (Renne et al., 1998). The Cochetopa Dome extruded within the Cochetopa Caldera; little palaeoenvironmental evidence has been presented for this eruption, but it’s eruption within an actively subsiding caldera filled with tuffs and volcaniclastic sediments (Lipman and McIntosh, 2008) suggests that this obsidian formed either subaerially, or perhaps in contact with caldera lake water.

Polished slabs were prepared from a piece of variably devitrified flow banded obsidian and from a 17 mm diameter fresh perlite bead selected from Sample OOL-31A (Fig. 3). The fresh perlite bead is transparent with faint flow bands displayed as subtle variations in colour and translucency. The variably devitrified slab displays a range of textures: flow banding in the freshest parts of the obsidian is seen as alternating light and dark bands of varying thickness and opacity running across the sample, related to variable number densities of microlites. Discontinuous bands of opaque, amorphous pink and grey material are interpreted as devitrified patches and are elongate sub-parallel to the flow bands in the fresh material. They represent spherulitic growth that originated from chains of band-parallel nuclei and are fringed by overgrowths displaying sharp, undulose and locally feather-like contacts with the surrounding glassy material. The glass is

![Image](https://example.com/fig3.png)

**Fig. 3.** Photographs of polished slabs of the large, fresh perlite bead and the variably devitrified flow banded obsidian. White lines show the fragments selected for UV-LAMP Ar-isotope analysis. On the fresh perlite bead, the dotted black lines show the positions of the innermost and outermost laser ablation trenches.
cut by a high density of concentric, curved perlitic cracks, which break up the fresh obsidian into small (1–7 mm) perlitic beads. The smallest beads are spatially associated with the most highly devitrified patches.

2.1. Ar-isotope analysis

300 µm-thick polished slabs were demounted and small (1 cm²) fragments from all the slabs were selected for Ar/Ar analysis, washed with acetone and deionised water, packaged in Al-foil, loaded into a foil tube and irradiated at the McMaster Reactor (Canada) for 16.7 h using cadmium shielding. The neutron flux was monitored using bismuth standard GA1550, grains of which were loaded into 1 cm foil packets and distributed throughout the irradiation tube. Grains from each packet were analysed by single grain fusion and J-values for each slab fragment were calculated by linear extrapolation of J-values between standards within the irradiation tube. J-values were calculated assuming an age of 98.79 ± 0.96 Ma (Renne et al., 1998) which is consistent with the fluence monitor age used by Lipman and McIntosh (2008) to allow direct comparison of our new ⁴⁰Ar/³⁹Ar data with theirs. A New Wave Research Ltd. LUP 213 nm, 20 mJ, pulsed Nd-YAG laser was used to ablate trenches across the polished surfaces of the slabs at 20 Hz and 65% power. The geometry of the trenches was texture-dependent and ranged from 0.3 mm² rasters to trenches 100 µm wide and up to 6 mm long. Lasering durations were typically 8 min. Significant redeposition of ablation ejecta was not observed via the optical camera attached to the laser system. Unwanted gas species were removed from released gases by two SAES getters, one operating at optical camera attached to the laser system. Unwanted gas species were removed from released gases by two SAES getters, one operating at room temperature and the other at 450 °C, for a minimum of 150 s prior to automatic inlet into a Nu Instruments Noblesse mass spectrometer, which is able to resolve ³⁶Ar from ³⁶(Cl). Masses 35, and interfering peak shoulders on the high mass sides of masses 36 (36-high) and 39 (39-high) were measured in addition to the Ar isotopes to monitor for chlorine and hydrocarbon interferences, respectively. No correlation between blank corrected ³⁶Ar and blank corrected ³⁶-hig or ³⁵Cl was observed (Fig. 4), suggesting that apparent ³⁶Ar abundances have not been artificially increased by contamination from hydrocarbon or ³⁶(HCl) mass interferences. Blanks were measured between each sample analysis. All data were corrected for blanks, ³⁷Ar decay and neutron-induced ⁴⁰Ar, ³⁹Ar and ³⁶Ar. Atmospheric ³⁹Ar calculations were carried out assuming an atmospheric argon composition of ⁴⁰Ar/³⁶Ar = 295.5 (Steiger and Jäger, 1977). All errors quoted in tables and text are 1 standard deviation unless otherwise stated. We note the existence of an update of the modelled value of atmospheric Ar (Lee et al., 2006) and for the age of GA1550 (Spell and McDougall, 2003), but have adopted the same external correction factors as those used by Lipman and McIntosh (2008) for ease of comparison between their data and ours. A minority of analyses produced ³⁵Ar and ³⁷Ar yields that were indistinguishable from blank and where this resulted in a negative blank-corrected figure, the value was assumed to be 0 for the purpose of Ca and atmospheric corrections to the ³⁹Ar/³⁹Ar age. ³⁷Ar yields were low for all analyses, which is not surprising given the low CaO content (0.41%) of the sample (Lipman and McIntosh, 2008). During Ar-isotope analysis, no unusual baseline fluctuations were observed and the peak positions for low-abundance isotopes (including ³⁶Ar) were monitored to ensure that the low measured abundances were real and not the result of measuring off-peak, so these exceptionally low abundances are considered to be real and not an artefact of ablating too little material or of decay of ³⁷Ar to undetectable levels.

2.2. Infra-red spectroscopic analysis

A doubly polished wafers of ~250 µm thickness was prepared from an equivalent section of the fresh perlitic bead. Unfortunately, pervasive fracturing of the devitrified sample defied all attempts to prepare doubly polished wafers. Infra-red spectroscopy was used to characterise dissolved water concentrations in the glass, using the system at Lancaster University (Thermo Nicolet IR interferometer with KBr beamsplitter, coupled to a Continuum Analytical microscope, with an MCT-A detector and 50 µm square aperture). 128 spectra were collected in the mid-IR range of 4000–1000 cm⁻¹. Twelve-point linear baseline corrections were used to quantify the H₂O absorption peak at 3550 cm⁻¹, and peak values were converted to species concentrations using the Beer-Lambert Law and a glass density of 2300 kg m⁻³. A Mitutoyo digital micrometer was used to measure wafer thickness, with an accuracy of ± 3 µm, and an absorption coefficient of 80 mol⁻¹ cm⁻¹ was used for the 3550 cm⁻¹ peak (Newman et al., 1986). The combined absolute uncertainty on such IR measurements is ~10%, with significantly smaller relative errors (e.g. von Aulock et al., 2014). The fresh perlitic bead was analysed by measuring the gas released from a series of 22 trenches excavated between the rim and the core of the bead. The analyses gave apparent Ar ages between 25.80 ± 0.57 and 26.99 ± 0.52 Ma. Eight of these analyses are within one standard deviation of Lipman and McIntosh's (2008) published age (26.91 ± 0.1 Ma) and all are within two standard deviations. There does not appear to be any systematic variation between the ages with distance from the bead rim. A weighted average of all 22 data points (excluding the error on the J-value) gives an age of 26.59 ± 0.06 (1 s.d.) with an MSWD (mean square of weighted deviates) of 0.37 and a probability of 0.996. An inverse isochron diagram of all data excluding the three analyses that did not yield ³⁶Ar above blank (n = 19) gives an age of 26.61 ± 0.17 Ma with a trapped ⁴⁰Ar/³⁶Ar of 283 ± 33, an MSWD of 0.41 and a probability of 0.98. The trapped ⁴⁰Ar/³⁶Ar value is indistinguishable from that of air and the large error on the intercept is due to clustering of the data at the radiogenic end of the isochron diagram. The weighted average and the isotropic age are indistinguishable and we adopt the more precise age of the weighted average. Recalculating the weighted average to include the error on the J-value gives an age of 26.58 ± 0.07 Ma. This age is 330 ka younger than the previously published age.

Distribution of K (as ³⁹Ar) and ³⁶Ar throughout the bead was investigated by normalising ³⁹Ar and ³⁶Ar to ³⁷Ar and plotting against rim-core distance. ³⁷Ar was chosen as the normalising factor as it is derived from Ca, which should remain immobile during hydration of the volcanic glass, although the low abundance of Ca in the sample results in relatively large errors on the ³⁷Ar values and subsequent

![Fig. 4. Apparent ³⁶Ar yields plotted against measured ³⁵Cl and ³⁶(Cl) to check for overestimation of ³⁶Ar due to contamination from species with an interference at mass-36. All values are blank corrected. There is no correlation between apparent ³⁶Ar yield and the potential contaminants, suggesting that the ³⁶Ar values are correct and have not been overestimated.](image-url)
ratios. 39Ar showed very little variation with distance from the bead rim, indicating a uniform K:Ca ratio throughout the perlite bead. 36Ar also appears uniform throughout the bead except for the outer few hundred microns where elevated 36Ar/37Ar values occur. These values are associated with higher errors; 37Ar analytical errors are proportionally larger in the first few analyses, due to smaller overall signal sizes while the optimum ablation length was being determined, and these errors propagate through to create larger errors on the isotope ratios. Elevated 36Ar is also reflected in significantly lower radiogenic 40Ar (40Ar*) yields in the outer few hundred microns (Fig. 5), as 40Ar* is calculated by subtracting atmospheric 40Ar (calculated based on measured 36Ar) from the total measured 40Ar.

3.2. Variably devitrified slab (fragment H)

Ar-isotopes were characterised for a number of discrete textural zones on the devitrified slab. The core-rim variation of three small perlite beads (A–C) was investigated by excavating concentric trenches in each bead. Flow banding in relatively fresh obsidian was investigated in a fourth small perlite bead (D) by rastering across areas of ~0.3 mm². Visually distinct bands and patches in the highly devitrified material were analysed by ablating trenches ~100 μm wide or rastering across areas up to 0.5 mm². The textural context and position of the analyses and a summary of the data as Ar/Ar ages is given in Fig. 6.

Highly devitrified material, appearing as flow bands and patches of opaque pink and grey material, gives a range of ages (24.09 ± 2.89 to 29.42 ± 1.45 Ma), most of which are slightly younger than the fresher material (Fig. 6). These patches yielded the highest 36Ar/37Ar values (0.14 ± 0.03 to 0.72 ± 0.55) and lowest 39Ar/37Ar values (5.7 ± 1.27 to 17.17 ± 9.47) of the sample, indicating loss of K and high levels of non-radiogenic Argon. An isotope correlation diagram of these phases gives an age of 25.6 ± 1.2 with a 40Ar/36Ar intercept of 298.6 ± 63, an MSWD of 2.2 and a probability of 0.086. The three small beads have diameters of 1.7 (bead A), 0.8 (bead B) and 1.4 (bead C) mm (Fig. 5). All three beads exhibit slight age gradients with age decreasing from rim to core. 39Ar/37Ar values show no systematic variation with distance from the bead rim, and are indistinguishable from the values produced by the fresh perlite bead, suggesting a uniform K-content (Fig. 7).

Bead A produced ages between 25.5 ± 0.37 and 26.26 ± 0.27 Ma. All of the ages are younger than the age of the fresh perlite bead, but are within three standard deviations of the weighted average. An inverse isochron plot (n = 5) gives an age of
Fig. 7. A: Core-rim age variation in the 3 perlite beads compared to the weighted average age of the fresh perlite bead and Lipman and McIntosh’s published age. B: Core-rim variation in 39Ar/37Ar as a proxy for K+Ca. Grey dashed line is the average value of the fresh perlite bead and is shown as a rough comparison only as the neutron fluence (i.e. 39Ar production) is slightly different between slabs – see Supplementary Data for J-values. C: Core-rim variation of 40Ar* yield compared to the outer 1 mm of the fresh perlite bead. D: Core-rim variation in 36Ar/37Ar compared to the average value of the inner portions of the fresh perlite bead (grey dashed line).

25.8 ± 2.4 Ma with a 40Ar/36Ar intercept of 300 ± 50, an MSWD of 1.0 and a probability of 0. 36Ar/37Ar values are higher than for the fresh perlite bead, with the highest values occurring in the outermost part of the bead. This elevated 36Ar is mirrored by relatively low 40Ar* yields (calculated based on measured total 40Ar and 36Ar) of ~70%.

Bead B produced ages between 26.52 ± 0.75 and 27.2 ± 0.4 Ma. All of the ages are within error of the weighted mean age of the fresh perlite bead but the analyses closer to the rim are older, while the core analyses are almost exactly the same as the fresh bead weighted average (Fig. 7). An inverse isochron plot gives an age of 26.67 ± 0.85 Ma with a 40Ar/36Ar intercept of 325 ± 68, an MSWD of 0.67 and a probability of 0.89. N = 4, but three of the data points are clustered and the trend of the isochron line is controlled by the high 36Ar analysis closest to the rim. Most of the 36Ar/37Ar values are comparable to that of the fresh bead apart from the analysis closest to the rim which exhibits a higher level of 36Ar and a corresponding lower 40Ar* yield (Fig. 7).

Bead C produced apparent ages between 26.11 ± 0.43 and 27.65 ± 0.3 Ma. This age range spans the weighted average of the fresh perlite bead but the outermost analysis is significantly older. An inverse isochron diagram shows that the Ar-isotope system in this bead is clearly disturbed as it gives an apparent age of 46 ± 140 Ma with a 40Ar/36Ar intercept of −373 ± 470, an MSWD of 0.67 and a probability of 0.51. 36Ar/37Ar values are higher than for the fresh perlite bead and 40Ar* yields are lower.

The fourth small bead (bead D, 2 mm diameter) was analysed to investigate any isotopic difference between dark and light (translucent) coloured flow bands. The light coloured flow bands were glass that appeared clear in thick section and relatively free from microlites and inclusions, while the dark flow bands appeared black in the thick section due to an abundance of microlites. Core-rim variation within each flow band was investigated by analysing 3 rastered trenches per flow band. Both flow bands give similar age ranges but the light flow band shows a progression of an older age (26.66 ± 0.35 Ma) at the rim to a younger age (25.62 ± 0.27 Ma) at the core, while the dark band does not show a systematic variation of age with core-rim distance (Fig. 8). 39Ar/37Ar values show no variation. The light flow band has a higher 36Ar content, with higher 36Ar/37Ar values showing a slight increase towards the core and lower 40Ar* yields (decreasing from 86% at the rim to 79% in the core) compared to the dark flow band. 36Ar yields are slightly higher (albeit within error due to low abundances of 36Ar and 37Ar) at the core compared to the rim in the dark flow band and the core-rim variation is much less than in the light flow band (Fig. 8).

4. Water concentration results

Water concentration data in an equivalent section through the fresh perlite bead are shown in Fig. 9, along with an image of the wafer in order to show the relative positions of the data points. The interior of the bead shows a remarkably homogenous water concentration, with mean of 0.22 ± 0.01 wt% (n = 102, 1 standard deviation), and range 0.20–0.23 wt%. By contrast, there is strong H2O enrichment within the outer 200 μm of one side of the wafer, reaching 0.68 wt% in the outermost analysis directly adjacent to the bead-bounding fracture. An equivalent enrichment could not be determined on the other side of the traverse due to an irregular wafer edge, which prevented reliable H2O determination.

5. Discussion

5.1. Obsidian texture and water content

The low water content of the interior of the fresh perlite bead is consistent with degassing to near-atmospheric conditions, and thus almost complete loss of the magmatic H2O component (e.g. Friedman and Smith, 1958). The elevated water content at the rim indicates strong but spatially-localised hydration, which is consistent with diffusion of H2O having advanced approximately 200 μm into the perlitic bead. This distance is consistent with other studies that have quantified H2O diffusion signatures in perlitised obsidian (Tuffen et al., 2010; von Aulock et al., 2013), albeit in far younger samples that have interacted with high-temperature aqueous fluids over far shorter timescales. We thus interpret this obsidian to have degassed the majority of its magmatic volatiles during eruption and to have undergone limited post-eruption hydration at the margins of perlite bead rims.
Water concentrations are not available for Fragment H, but the presence of both perlitic textures and strongly devitrified bands indicates the involvement of water mobility in the alteration process. It is well established that perlitised obsidian displays elevated H2O contents that can be resorbed by surrounding obsidian (Castro et al., 2008; Gardner et al., 2012; von Aulock et al., 2013), with D/H ratios indicating ingress of meteoric water (Bindeman and Lowenstern, 2016). Additionally, the growth of anhydrous mineral assemblages during devitrification (Manley and Fink, 1987), creates diffusive fronts of magmatic volatiles that flux through the surrounding obsidian (Castro et al., 2008; Gardner et al., 2012; von Aulock et al., 2013). The latter process can therefore liberate and concentrate a magmatic H2O component that can then be resorbed during obsidian cooling and hydration. This conflicts with the traditional model that hydration involves uniquely meteoric water (e.g. Ross and Smith, 1955). Indeed, models of the textural evolution of obsidian flows describe how crystallisation of anhydrous phases within flow interiors liberates magmatic H2O, which migrates and resorbs into the upper part of the flow, driving vesiculation (Manley and Fink, 1987). These vesicles may, in turn, act as nucleation points for spherulite growth (Seaman, 2013), which then promotes further magma H2O mobility. Increased H2O solubility during cooling of obsidian promotes retrograde water resorption and enrichment, with H2O derived from either magmatic or meteoric sources (Ryan et al., 2015). The textures in Fragment H thus reflect the higher volatile content of this obsidian than the fresh bead, with volatile enrichment either reflecting ingress of magmatic volatiles locally liberated by devitrification, absorption of meteoric water and other volatiles due to retrograde solubility during cooling, less efficient degassing of magmatic volatiles, or any combination of these three mechanisms.

5.2. Spatial variation in Ar-isotopes

Ar-isotopes show significant spatial variation within and between different obsidian textures and microtextures. In many cases, these observations are not consistent with simple binary mixing between radiogenic and a trapped end-members – a condition necessary for robust 40Ar/39Ar dating. Key observations of Ar-isotope distribution are:

1. The low 36Ar content (using 36Ar/37Ar as a proxy) of the fresh perlite bead interior and Bead B of Fragment H (Figs. 5 and 7).
2. The higher 36Ar content at the rim of the fresh perlite bead and Bead B of Fragment H (Figs. 5 and 7).
3. Lower 40Ar* yields in the outer few hundred microns of the fresh bead and Bead B (Fig. 7).
4. Decreasing apparent 40Ar/39Ar age from rim to core in the perlite beads of Fragment H (Figs. 6 to 8).
5. The difference in 36Ar content between the dark and pale flow bands in Bead D of Fragment H (Fig. 8).
6. The decrease in 40Ar* yield from rim to core in the pale flow band of Bead D in Fragment H (Fig. 8).
7. The increase in 36Ar content (using 36Ar/37Ar as a proxy) from rim to core in the pale flow band of Bead D (Fig. 8).
8. The range in different proportions of radiogenic to trapped argon...
Fig. 10. Inverse isochron diagram showing the relationship of the various obsidian microtextures and the expected data trajectories that would be produced by changes to the Ar-isotope system. Inset shows the full range of data points with the isochron line for the fresh perlite bead. The boxed area on the inset shows the extent of the main diagram, which is expanded to include more detail while excluding four of the devitrified analyses. The thick, solid line is the isochron for the large fresh bead, dashed line is the isochron for Bead A, dotted line is the isochron for Bead B and thin solid line is the isochron for bead C.

between perlite beads in Fragment H (Fig. 10).

Below, we discuss possible mechanisms that could generate these observations, in the context of the textural and volatile interpretations above, to derive hypotheses for the volatile and Ar-isotope evolution of these samples.

As previously discussed, mixing between radiogenic argon and multiple reservoirs of trapped argon (e.g. an atmospheric and an excess \(^{40}\)Ar component, or multiple excess \(^{40}\)Ar components) have the potential to produce statistically robust, but meaningless isochrons (Flude and Storey, 2016; Kuiper, 2002; Sherlock and Arnaud, 1999). Fig. 10 plots the obsidian data on an isotope correlation diagram and shows the expected trajectories that data points would follow if the Ar-isotope system were disturbed by loss of \(^{40}\)Ar, loss of K, addition of atmospheric argon, or addition of excess \(^{40}\)Ar. Much of the spread in data between different obsidian textures can be superficially explained by mixing between a radiogenic end-member and different proportions of trapped argon with a roughly atmospheric composition. The fresh perlite bead plots at the radiogenic end of the isotope correlation diagram, along with Bead B and the dark flow band from Bead D. The least radiogenic data come from the strongly devitrified areas.

5.3. Argon reservoirs

In the case of the fresh perlite bead and Bead B, much of the Ar-isotope data appears to be radiogenic (Fig. 10). The variation in Ar-isotopes correlates with water content, with higher proportions of \(^{36}\)Ar and higher water concentrations occurring in the outer 400 and 200 μm, respectively. The low proportion of non-radiogenic argon in the interior of the fresh bead is consistent with the interpretation, based on water concentration, that this sample is well degassed from its magmatic volatiles. The coincidence of elevated water and \(^{36}\)Ar content on water concentration, that this sample is well degassed from its

5.4. Obsidian argon degassing profiles

All of the perlite beads in Fragment H show variation in apparent \(^{40}\)Ar/\(^{39}\)Ar age, with ages decreasing from rim to core. This age pattern cannot be explained by diffusive loss of \(^{40}\)Ar* from the bead rims over geological time, as this would produce younger rims and older cores. Older ages at bead rims are consistent with diffusive uptake of excess \(^{40}\)Ar during interaction with an external excess \(^{40}\)Ar rich-fluid or melt (Kelley, 2002). Flushing of excess \(^{40}\)Ar-bearing gas through the lava may occur during eruption close to the vent area, but this is likely to occur at near-atmospheric pressure. As discussed above, argon solubility at low pressures in silicate melts and glasses is low, therefore high concentrations of argon would need to be present in the gas phase for this to partition into the obsidian via simple diffusion. Carbon dioxide accumulations at Sheep Mountain and McElmo Dome, located east and west of the San Juan Mountains, respectively, are derived from magmatism associated with the Colorado Plateau uplift and contain maximum concentrations of 0.02% and 0.33% \(^{40}\)Ar, respectively (Gilfillan et al., 2008). These are comparable to \(^{40}\)Ar concentrations of up to 0.37% in volcanic gases escaping from the base of Mount Etna (Pagonita et al., 2012). Concentrations of argon in magmatic gas are thus much lower than the atmospheric concentration of argon (0.9%, Ozima and Podosek, 2002) and so post-eruptive uptake of excess \(^{40}\)Ar by obsidian from magmatic gas, through simple diffusion, is unlikely to significantly increase dissolved \(^{40}\)Ar concentrations. However, hydration of the obsidian by a fluid containing excess \(^{40}\)Ar, as may happen if the hydrating fluid is magmatic or has been interacting with old, K-rich continental crust, may provide a mechanism to incorporate excess \(^{40}\)Ar into the perlite bead rims. In this scenario, we would expect a spike in \(^{40}\)Ar/\(^{39}\)Ar age and \(^{40}\)Ar+% at the hydrated rim of the bead, with consistent values throughout the bead interior, or a more gentle decrease and plateau if temperatures are high enough to allow diffusion of Ar through the obsidian silicate network after hydration. Instead, gentle \(^{40}\)Ar/\(^{39}\)Ar age gradients are observed throughout the small perlite beads, from core to rim, and \(^{40}\)Ar yields are either consistent across the bead, or show a decrease at the rim (Bead B, Fig. 7), suggesting addition of atmospheric argon during hydration. The only data set that shows a decrease in \(^{40}\)Ar yield from rim to core is from the pale flow band of Bead D, but this correlates with an increase from rim to core in \(^{36}\)Ar yield (Fig. 8).
This variation in $^{36}$Ar yield is best explained by loss of $^{36}$Ar from the bead margins; it is difficult to envisage a scenario where $^{36}$Ar is preferentially added to the core of a bead. This would suggest that the $^{36}$Ar data represents a degassing profile that is not observed in the dark flow band because this has already more efficiently degassed. If such degassing took place without isotope fractionation, the result would simply be variable proportions of nonradiogenic argon and, assuming the trapped argon isotopic composition is well known, all analyses should give consistent ages. However, this is not observed for the pale flow band in Bead D, or for the other perlite beads in Fragment H.

As previously discussed, Ar-isotope fractionation during diffusion has been cited as a possible mechanism for producing age variation in obsidian and specifically for the low (< 295.5) $^{40}$Ar/$^{36}$Ar trapped Ar compositions that have been determined by various authors (Brown et al., 2009; Morgan et al., 2009; Vogel et al., 2006). In these scenarios, incomplete equilibration of the obsidian with Ar-bearing air/water is suggested as way of increasing the $^{36}$Ar/$^{40}$Ar, ratio, as diffusion of lighter $^{36}$Ar will outstrip that of $^{40}$Ar, resulting in excess $^{36}$Ar. These studies have focussed on Ar gain by cooling obsidian, but have not considered similar fractionation during diffusive Ar loss during degassing. In such a scenario, the faster rate of $^{36}$Ar diffusion may lead to preferential $^{36}$Ar loss, resulting in a trapped Ar composition of $^{40}$Ar/$^{36}$Ar > 295.5. This mechanism may thus be the cause of the excess $^{36}$Ar found in some obsidians (e.g. Flude et al., 2010).

Better quantification of the different diffusivities of Ar-isotopes in silicate networks is required before such fractionation processes can be quantitatively modelled. Nonetheless, it is useful to qualitatively assess our data in terms of these fractionation models. The intercept values of the isochrons for the perlite beads all fall within error of the atmospheric Ar value, but often because the error on the intercept is relatively large, due to either relatively poor correlations or limited spread of data. As such, precise values for the composition of the trapped Ar in each perlitic bead are unavailable and cannot be used to assess the presence of fractionation. However, a qualitative assessment can be made in that, when applying an atmospheric correction for trapped $^{40}$Ar, excess $^{36}$Ar will result in over-correction, resulting in younger ages, while excess $^{36}$Ar will result in under-correction and give older ages. Beads A–D all exhibit apparently older rims and younger cores, in-keeping with our diffusive fractionation hypothesis. To further test this we plot isotope correlation diagrams for core (innermost ablation trenches – Figs. 3 and 6) and rim (outermost ablation trenches – Figs. 3 and 6) analyses from beads A–D (Fig. 11). If our hypothesis is correct we would expect the rim analyses to give an intercept indicative of excess $^{40}$Ar, while the core analyses give an intercept of atmospheric composition.

An isochron of the core analyses gives an Ar/Ar age of 26.54 ± 0.64 Ma, a $^{40}$Ar/$^{36}$Ar intercept of 268 ± 24, and MSWD of 0.73 and a probability of 0.53. The rim data isochron gives an Ar/Ar age of 26.5 ± 2.2 Ma, a $^{40}$Ar/$^{36}$Ar intercept of 309 ± 83, an MSWD of 4.9 and a probability of 0.002 (Fig. 11). Unfortunately the poor correlations produced by scatter in the data precludes calculation of precise trapped Ar compositions. This is not surprising given the complex interplay of degassing and absorption experienced by different sized beads, all of which were presumably acting as individual domains and experienced different degrees of Ar loss and gain during degassing and absorption. Furthermore the imprecise geometric control on the plane of section of the beads inhibits precise quantitative modelling. Both intercept values are indistinguishable from the atmospheric value and each other at two standard deviations, with the poorest correlation (thus highest errors) being produced by the rim analyses which were likely subject to the most variable conditions during and after cooling of the obsidian. Nevertheless, it is clear from the isochron correlation diagram that the rim analyses trend towards a lower $^{38}$Ar/$^{40}$Ar (i.e. higher $^{40}$Ar/$^{36}$Ar) value than the core analyses and these observations are consistent with our hypothesis.

If the observed apparent $^{40}$Ar/$^{39}$Ar age gradients are the result of preferential loss of $^{36}$Ar during degassing, leading to a relative excess of $^{40}$Ar at the bead rims, we might expect gradients in $^{36}$Ar to be present in all of the affected beads, but this is only observed in Bead D. $^{36}$Ar/Ar ratios in Beads A and C are approximately one order of magnitude higher than observed for Bead D (Figs. 7 and 8), while the rim of Bead B has a similar $^{36}$Ar/$^{40}$Ar value to Beads A and C, but the interior is more similar to that of Bead D. This enrichment in $^{36}$Ar at the rim of Bead B is also reflected in a lower $^{40}$Ar* yield at the rim, and has been interpreted as an overprinting with atmospheric argon during hydration. The lack of observed $^{36}$Ar enrichment at the rim of bead D may be a sampling artefact related to the geometry of the raster pits. The elevated $^{36}$Ar (and corresponding lower $^{40}$Ar* yields) in Beads A and C could either be due to a high initial volatile content (due to proximity to the devitrified area, postulated to expel volatiles during crystallisation), or due to later addition of argon during hydration. If the Ar-isotope composition of the perlite beads was overprinted with atmospheric argon during late-stage hydration, this would likely mask any gradients in $^{36}$Ar that formed during degassing. However, adding unfractionated atmospheric argon would dilute, but not alter the apparent excess $^{40}$Ar resulting from diffusive fractionation, and the apparent $^{40}$Ar/$^{36}$Ar age gradients would remain.

5.5. Volcanological interpretation and lessons for $^{40}$Ar/$^{39}$Ar sampling

The Ar-isotopes in this obsidian thus appear to have derived from 3 possible sources. The first is initial Ar dissolved in the melt and that remained trapped in the obsidian as it chilled. The bulk isotopic composition of this component will depend on the nature and setting of the magmatic system but can be considered to have a $^{40}$Ar/$^{36}$Ar ratio of 295.5 or higher (i.e. of atmospheric composition, or containing excess $^{40}$Ar). This is generally assumed to be magmatic Ar, although recent observations of explosive activity during obsidian formation open up the possibility of bulk-incorporation of atmospheric Ar into extruded melt via repeated cycles of fragmentation, adsorption and rehealing. Such fragmentation and healing cycles occur at eruptive temperatures (Tuffen et al., 2003) and lead to the transient formation of permeable pathways connecting extruding magma with the surrounding hydrothermal system (Saubin et al., 2016). We thus refer to this component, which we consider to be present in the melt phase within subaerially emplaced lava, as "initial Ar". Poorly degassed obsidians, such as those explosively ejected from depth in the conduit as pyroclasts, should have higher trapped initial Ar contents than well-degassed melt typical of subaerial lava flows. As discussed above, this initial argon may be redistributed and concentrated within an obsidian flow by volatile expulsion from the crystallising and devitrifying flow interior. The second source is Ar from atmospheric sources incorporated.
into the obsidian during or after quenching via diffusion, hydration or other alteration mechanisms; we consider this to have a bulk reservoir composition similar to that of atmospheric Ar. The third source is radiogenic $^{40}$Ar* derived from the radioactive decay of $^{40}$K over geological timescales.

To deduce accurate $^{40}$Ar/$^{36}$Ar and K-Ar ages the contribution of the first and second of these components needs to be deconvolved from the radiogenic $^{40}$Ar*. This is usually done by assuming that all non-radiogenic Ar has an atmospheric composition, and thus correcting the $^{40}$Ar value based on the $^{36}$Ar yield. Any $^{36}$Ar present in the obsidian is non-radiogenic, so must be derived from a trapped initial component or a post-eruption meteoric component, and so the distribution of $^{36}$Ar within the obsidian may give insights into the origin of the non-radiogenic Ar and the degassing and/or absorption processes which took place in the obsidian.

The large, fresh perlite bead produced consistent $^{40}$Ar/$^{36}$Ar ages and displayed a consistently low $^{36}$Ar content, apart from at the rim where $^{36}$Ar is elevated (Fig. 5). As discussed, this elevated $^{36}$Ar at the bead rim is thought to have been introduced via hydration. The increase in $^{36}$Ar without a corresponding change in apparent Ar-age suggests that non-fractionated atmospheric Ar was incorporated during hydration. The otherwise low $^{36}$Ar content suggests that any initial Ar in the melt was efficiently degassed in this sample, consistent with uniformly low H$_2$O concentrations far from the bead margins.

In the case of perlite Bead D, the dark, microlite rich flow band gave a lower $^{36}$Ar yield than the light flow band. Given the lack of alteration in the light flow band, this is consistent with a greater degree of degassing of initial Ar in the dark flow band, enhanced by and/or promoting the formation of microlites. Both flow bands exhibit lower $^{36}$Ar yields towards the rim of the bead, again consistent with degassing of magmatic Ar during cooling and indicating that this perlite bead formed during an early post-eruptive phase with elevated temperatures and diffusively mobile magmatic Ar, rather than during a later devitrification event. Overprinting with atmospheric argon during hydration at the rims of Beads B and the fresh bead, and the postulated more diffuse overprinting of beads A and C indicate that a second stage of hydration took place at a later post-eruptive stage. Contrasting degassing and absorption histories appear to be present within just a few square millimetres, and these go some way to explaining the difficulties with accurately interpreting $^{40}$Ar/$^{36}$Ar data from obsidians.

Fig. 12 summarises our proposed model for argon isotope evolution in Cochetopa Dome obsidian, and is used to illustrate our recommendations for selecting obsidian samples for the most reliable $^{40}$Ar/$^{36}$Ar dating. We simplify argon-isotope evolution into three stages each representing the action of one of the three sources of argon, described above.

In Stage 1, the initial argon content of the melt is distributed throughout the obsidian; some parts undergo effective degassing while other areas may be enriched due to fluxing of volatiles from crystallising or devitrifying parts of the lava (Fig. 12B). Perlitic cracks form while the lava remains at high enough temperatures ($> 500 \, ^\circ \text{C}$) to promote argon diffusion through the glass structure (Fig. 12C). The perlitic cracks act as diffusion domain boundaries and argon diffuses out of the obsidian, being lost to the fast diffusion pathways at the bead rims. Degassing produces concentration gradients with depleted bead rims (Fig. 12D). Kinetic fractionation during diffusive degassing results in variable $^{40}$Ar/$^{36}$Ar values across the bead with an apparent excess of $^{40}$Ar (compared to air) at the rims due to preferential loss of $^{36}$Ar during degassing. At the end of Stage 1, argon concentrations and isotope ratios are variable throughout the sample due to multiple degassing and enrichment pathways and variable degrees of kinetic isotope fractionation.

During Stage 2, the obsidian undergoes hydration; this introduces atmospheric argon, which overprints the initial argon isotope and concentration gradients (Fig. 12E). This may occur during the final stages of eruption, or over millions of years. Stage 3 represents the time period between chilling of the lava below $\sim 500 \, ^\circ \text{C}$ and sample collection (in this case, $\sim 27 \, \text{Ma}$). $^{40}$K radioactively decays to $^{40}$Ar* (Fig. 12F).

Such obsidians pose a problem for K-Ar and $^{40}$Ar/$^{36}$Ar dating because of the heterogeneous initial argon content (especially in terms of isotope ratios) and because addition of atmospheric argon during hydration results in mixing of three sources of argon, making it impossible to characterise the initial component. As such, parts of obsidian that are well degassed from their initial argon, and have not experienced secondary hydration are most likely to produce reliable K-Ar and $^{40}$Ar/$^{36}$Ar ages. In Fig. 12, these areas correspond to the yellow areas of panel F, and green (hydrated) and orange (poorly degassed) areas should be avoided when sampling for Ar-dating (see online version of the article for a full colour version of the figure).

### 5.6. Implications for volcanological study of obsidian

Quantification of Ar systematics at high spatial resolution within obsidian, when combined with their textural context within rhyolitic lava, can shed light on late-stage fluid flow and its association with crystallisation. Patterns of $^{36}$Ar within small perlitic beads A-D from the partly-devitrified zone are consistent with diffusive $^{36}$Ar loss at comparatively high temperatures, in contrast with the inferred lower-temperature Ar diffusion at the margin of the larger bead within fresher, non-devitrified obsidian. High-temperature fracturing of melt/glass adjacent to spherulitic masses, as seen in rhyolitic obsidian from Krafla, Iceland (Castro et al., 2008), may be driven by volumetric changes upon crystal growth, together with localised magmatic fluid overpressure. These fractures provide pathways for transport of magmatic gases that have been locally exolved around spherulites, due to enrichment of volatiles at the boundary of the growing anhydrous crystalline assemblage.

The occurrence of fracture-driven volatile mobility associated with zones of high-temperature devitrification supports a key hypothesis about the emplacement of rhyolitic obsidian flows. The formation of late-stage pumice diapirs, prominent on the surface of many of the Holocene silicic flows in the western US (e.g. Fink, 1983), has been attributed to the release of volatiles during crystallisation of the slowly-cooling flow interior, and their transport within ephemeral fracture networks (Manley and Fink, 1987). However, spherulitic textures within US flows are only clearly found within core samples, and little evidence has been presented to date for the nature of the relic pathways that facilitated gas transport away from zones of spherulite crystallisation.

### 6. Implications and conclusions

UV-LAMP Ar-isotope data from a range of obsidian microtextures within a single sample, are consistent with the presence of three separate reservoirs of argon in the obsidian (initial argon, atmospheric argon introduced during hydration, and $^{40}$Ar*), which were incorporated and modified during different phases of the emplacement and initial cooling of obsidian lava, together with later low-temperature, post-eruptive processes. Deconvolving these sources and processes can aid interpretation of apparent Ar-ages, and the physico-chemical evolution of rhyolitic lava.

The initial argon component of a melt (which may be magmatic, or atmospheric and incorporated during fragmentation and reheating on eruption) may be modified by heterogeneous degassing and volatile enrichment throughout the lava, both above and below the glass transition, with isotope fractionation occurring during partial degassing. This heterogeneous distribution of Ar-isotopes is retained once the glass cools below the argon closure temperature (measurable Ar-loss would take years at temperatures below $\sim 500 \, ^\circ \text{C}$). Atmospheric argon is subsequently introduced to parts of the obsidian during hydration, which may happen during and immediately after the eruption, or over
the millions of years following eruption until the lava has completely devitrified and no glass remains. Deconvolving these two sources of non-radiogenic argon from $^{40}$Ar$^*$ is essential for reliable K-Ar and $^{40}$Ar/$^{39}$Ar dating, but is difficult. The resulting obsidian sample contains a mixture of variably modified argon reservoirs, making isotope correlation diagrams difficult to interpret in terms of eruption age. Initial argon may fractionate during diffusive degassing, with preferential loss of $^{36}$Ar, resulting in an apparent excess of $^{40}$Ar at diffusion domain rims. These rims of excess $^{40}$Ar may be subsequently overprinted by atmospheric Ar during obsidian hydration, making identification of the excess $^{40}$Ar impossible to quantitatively identify.

The observed age gradients in the small perlite beads are consistent with degassing profiles of around 500 μm in length (Fig. 6). This length scale is consistent with likely widths of diffusion zones at realistic obsidian cooling rates (Fig. 2).

This study has found no evidence for excess $^{36}$Ar, as proposed in other Ar-isotope studies on obsidian (Brown et al., 2009; Morgan et al., 2009), although it is possible that more detailed work on perlite bead rims may reveal a preferential gain of atmospheric $^{36}$Ar during hydration. Considering argon solubilities and diffusivities in silicate melts and timescales of obsidian formation, previous hypotheses regarding kinetic fractionation during diffusion of atmospheric argon into obsidian lava flows seem unlikely. However, our data supports models in which kinetic fractionation of Ar-isotopes can influence $^{40}$Ar/$^{39}$Ar apparent ages. Contamination in the mass spectrometer at mass 36 has yet to be ruled out, but if this phenomenon is real, a more likely explanation is that atmospheric argon becomes incorporated into the obsidian melt during fragmentation in the vent (Castro et al., 2014; Tuffen et al., 2003), possibly due to a combination of diffusion and surface adsorption, both processes having the potential to induce mass fractionation (van Zuilen et al., 2016).

Our study used one of the oldest obsidian samples available, in the hope that the age would result in high enough $^{40}$Ar$^*$ yields to provide precise dates from the very small volumes sampled during UV-LAMP analysis. A complimentary experiment will be to analyse zero-age obsidian, to characterise the distribution and isotopic composition of initial argon, without overprinting by $^{40}$Ar$^*$. Two Chilean volcanoes have erupted rhyolitic obsidian in the last 10 years (Chaitén and Cordén Caulle) and characterisation of Ar-isotope distribution in samples from these eruptions is a logical next step towards testing the hypotheses presented in this study.

When considered in terms of pre-eruption melt storage, subsurface melt degassing may promote preferential loss of $^{36}$Ar and thus a melt enriched in excess $^{40}$Ar. This may explain how obsidians can contain excess $^{40}$Ar, despite being stored and erupted through young, low-K crust (e.g. at Kerlingarfjöll - Flude et al., 2010). More detailed studies on the trapped Ar content of phenocrysts and volcanic glass may thus provide a useful tool to study magma storage and eruption processes.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.chemgeo.2017.05.018.