Micro-analytical Perspectives on the Bishop Tuff and its Magma Chamber

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ABSTRACT

New in-situ major and trace element analytical data from crystals (sanidine, plagioclase, biotite, orthopyroxene, clinopyroxene) and matrix glasses are presented from juvenile materials representing the full Bishop Tuff sequence from the earliest fall unit (F1) to the latest ignimbrite package (Ig2Nc). These data are combined with published information to investigate the nature and zonation of the pre-eruptive Bishop magma chamber. Our data confirm that this magma chamber was a single unitary body which was thermally and compositionally zoned. The zonation was largely established prior to the growth of crystals, and also prior to mixing in the lower parts of the chamber induced by late-stage intrusion of a magma of contrasting composition and slightly hotter temperature (the ‘bright-rim’ magma). Sparse mixed swirly and dacitic pumices show enrichments in Ba, Sr and Ti that identify these pumices as possible representatives of the ‘bright-rim’ magma. A model (revised from previously published work) for the pre-eruptive magma chamber comprises three main parts: (i) an upper, volumetrically dominant (~2/3), relatively unzoned region which was the source of the earlier, eastern-erupted ignimbrite units and their coeval fall units; (ii) a volumetrically minor transition zone which shows evidence for minor degrees of mixing and was the dominant source for the latest, eastern-erupted part of Ig1Eb (Sherwin subunit) and the earlier part of the northern-erupted ignimbrite (Ig 2Na); and (iii) a lower, volumetrically subordinate (~1/3) region which was affected by mixing with the ‘bright-rim’ invasive magma in the lead-up to the eruption, and fed later northern-erupted units. Ingress of the ‘bright-rim’ magma introduced orthopyroxene and bright-rimmed zircon crystals, induced partial resorption then overgrowth of rims enriched in Ti, Sr and Ba on sanidine and quartz, and development of zoning in clinopyroxene. Based on pumice proportions and associated crystal and glass chemistries through the eruptive sequence, we infer that the roof and floor of the magma chamber were stepped down to the north, such that the transition
zone magma formed the floor of the southern part of the melt-dominant chamber and the roof of the northern part. Our data reinforce the previous concept of a single compositionally and thermally zoned Bishop magma chamber and additionally provides a temporally constrained role for pre-eruptive magma mixing and the introduction of melts and minerals with contrasting compositions to the resident Bishop magma.

INTRODUCTION

Supereruptions (those involving $>10^{15}$ kg or $\sim$450 km$^3$ of magma: Sparks et al., 2005; Self, 2006; Miller & Wark, 2008) are the largest explosive volcanic eruptions on Earth, with both devastating immediate impacts and significant long-term environmental consequences (e.g. Rampino & Self, 1993; Sparks et al., 2005; Self, 2006; Self & Blake, 2008). Because of the comparative scarcity of these events, it is necessary to examine the geological record in order to understand the processes by which these giant magma bodies are assembled and stored in the crust, and how their eruptions are triggered.

The Bishop Tuff is an iconic example of a supereruption deposit formed from the eruption of $>600$ km$^3$ of compositionally- and thermally-zoned silicic magma (Hildreth & Wilson, 2007), and has been a focal point for $>35$ years of reconstructions of large magma chambers. The wealth of analytical investigation that has been undertaken on the Bishop Tuff (including, but not limited to: Hildreth, 1977, 1979; Halliday et al., 1984; Dunbar & Hervig, 1992a; Hervig & Dunbar, 1992; Christensen & DePaolo, 1993; Christensen & Halliday, 1996; Davies & Halliday 1998; Wallace et al., 1999; Anderson et al., 2000; Reid & Coath, 2000; Bindeman & Valley, 2002; Simon & Reid, 2005; Hildreth & Wilson, 2007; Simon et al., 2007; Wark et al., 2007; Reid et al., 2011; Gualda et al., 2012a; Pamukcu et al., 2012; Roberge et al., 2013; Chamberlain et al., 2014a; Simon et al., 2014) has led to it being used as a case study to build and refine models applied to the origins of large-volume, high-silica
rhyolitic volcanism. These models not only include those used to investigate the generation and evacuation of these supereruptive units (including the magma ‘mush model’; e.g. Bachmann & Bergantz, 2004; Gualda et al., 2012a; Carrichi et al., 2014; Malfait et al., 2014) but also the calibration of geothermometers such as Ti-in-quartz and Fe–Ti-oxide (e.g. Wark et al., 2007; Thomas et al., 2010; Thomas & Watson, 2012; Wilson et al., 2012; Evans & Bachmann, 2013; Ghiorso & Gualda, 2013; Evans et al., 2015).

Although extensively studied, there is currently debate around the Bishop magma system, centred around the mineral (rather than glass and melt inclusion) records, with two contrasting approaches and conclusions.

(1). As originally proposed by Hildreth (1977, 1979), the Bishop Tuff was erupted from a unitary body of magma that was compositionally- and thermally-zoned prior to crystal growth. This hypothesis was established using comprehensive petrography and analysis of all crystal phases in the Bishop Tuff. The variations in crystal composition (interpreted to reflect growth of crystals in the melt in which they were found) thus were taken to reflect compositional stratification of the Bishop magma chamber, prior to crystallisation, from upper, more-evolved, cooler to lower, less-evolved, hotter magma (Hildreth, 1977, 1979; Hildreth & Wilson, 2007). This hypothesis has been strengthened by further whole-rock data (e.g. Hildreth & Wilson, 2007), and melt inclusion studies which show that the upper-regions of the magma chamber were richer in water, and poorer in CO₂ when compared with the lower, less-evolved, water-poorer, CO₂-richer regions of the Bishop magma chamber (Dunbar & Hervig, 1992a; Wallace et al., 1995, 1999; Anderson et al., 2000; Roberge et al., 2013). This compositional gradient is mirrored by an inferred thermal gradient of 80-100 °C, with coolest temperatures (modelled from Fe-Ti oxide, δ¹⁸O (Qtz-mt), and two-feldspar thermometry: Hildreth 1977, 1979; Bindeman & Valley, 2002; Hildreth & Wilson, 2007; Ghiorso & Evans, 2008; Chamberlain et al., 2014b) being at the top of the magma chamber.
(2). Building on suggestions that either or both of the Fe-Ti oxides and pyroxenes are not in equilibrium with the melt (Andersen & Lindsley, 1988; Lindsley et al., 1991), thermodynamic modelling has been used to suggest that values for pre-eruptive temperature gradients are flawed (Ghiorso & Gualda, 2013). Associated Rhyolite-MELTS modelling and silica-in-glass barometry of the Bishop Tuff has been used to assert that there was no significant thermal or pressure gradient within the Bishop magma chamber, with all samples [from the Hildreth (1977) dataset] displaying apparent pressures and temperatures of 175 – 200 MPa and 740 – 750º C, respectively (Gualda et al., 2012a; Gualda & Ghiorso 2013a,b).

This similarity of pressure and temperature estimates combined with a perceived bimodality of crystal compositions, has led to the hypothesis that the Bishop Tuff is the product of the eruption of two chemically distinct and physically separated magma bodies (Gualda & Ghiorso, 2013a), with analogues drawn from the multiple magma bodies of the Kidnappers (Cooper et al., 2012) and Mamaku/Ohakuri (Gravley et al., 2007) eruptions in New Zealand.

The dependence of many models for silicic magmatism on data and concepts from the Bishop Tuff means that developing an accurate understanding of the nature and stratification (both thermal and compositional) of the pre-eruptive Bishop magma body is of great importance. Debates regarding the nature of the Bishop Tuff magma body (or bodies) have, however, relied solely on the original compositional data for pumices and crystals (as compared to melt inclusions) of Hildreth (1977, 1979)(e.g. Evans & Bachmann, 2013; Gualda & Ghiorso, 2013a). Only Fe-Ti-oxide compositional analyses and bulk rock analyses have been carried out since (Hildreth & Wilson, 2007). In order to properly address the issues raised above, we present detailed in-situ analyses of crystals and glasses from the Bishop Tuff, that integrate textural and trace element compositional variations at the scale of tens of microns. We combine our spatially-resolved crystal compositional data with the eruptive stratigraphy of Wilson & Hildreth (1997) and Hildreth & Wilson (2007) to:
(a) re-address the composition of and stratification within the Bishop Tuff magma body,
(b) decipher the nature and role of the contrasting magma which gave rise to compositionally contrasting rims on sanidine, quartz and zircon (Wark et al., 2007; Gualda et al., 2012b; Chamberlain et al., 2014a,b), and
(c) investigate the role of subordinate, poorly preserved ignimbrite packages (Ig1SW, Ig1NW, Ig2SW; Fig. 2) in constraining any variability in the Bishop magma chamber.

**THE BISHOP TUFF**

The Bishop Tuff was erupted at 767.4 ± 2.2 ka (Rivera et al., 2011), accompanying the formation of Long Valley caldera in eastern California (Fig. 1: Bailey et al., 1976; Hildreth, 1979, 2004; Bailey, 1989; Wilson & Hildreth, 1997; Hildreth & Wilson, 2007). The >600 km³ of high-silica rhyolite magma is inferred to have been erupted, with possibly one short hiatus, over a period of ~6 days (Wilson & Hildreth, 1997). The Bishop Tuff represents the culmination of 4.5 Myr of magmatism in general and >1.4 Myr of rhyolitic volcanism in particular in the Long Valley region (Bailey et al., 1976; Metz & Mahood, 1985, 1991; Metz & Bailey, 1993; Bailey, 2004; Hildreth, 2004; Sarna-Wojcicki et al., 2005). In its proximal outcrop area between the Sierra Nevada and the White Mountains (Fig. 1) the tuff consists of nine fall units (F1-F9, Fig. 2), now preserved within a 90° sector east of the caldera, and multiple largely coeval ignimbrite packages (Ig1Ea to Ig2Nc: Fig. 2) mostly deposited in two sectors to the north and east of the caldera margin (Fig. 1). Lithic assemblages in the tuff link earlier erupted units to a vent area sited in the south-eastern part of what became the caldera, and reflect in their changes the subsequent concurrent development of venting along the northern and eastern portions of the caldera ring fracture (Figs. 1, 2: Hildreth & Mahood, 1986; Wilson & Hildreth, 1997).
Through the Bishop Tuff, the types and proportions of pumice change systematically (Hildreth & Wilson, 2007), in general from the dominance of crystal-poor (xp; <6 wt%) pumice in the earlier erupted units (F1-F8; Ig1 packages deposited mostly east of source) through crystal-moderate (xm; 6-12 wt %) to crystal-rich (xr; >12 wt %) pumice in the later erupted ignimbrite, deposited mostly north of the contemporaneous caldera rim. Petrographic studies (Hildreth, 1977, 1979) showed that sanidine and quartz are the most common crystal species present in the Bishop Tuff, found in roughly equal proportions. Plagioclase crystals commonly make up ~10% of the crystal assemblage. When combined, sanidine, quartz and plagioclase make up 98 – 99% of the total crystal population of the Bishop Tuff. Minor proportions of titanomagnetite and biotite are ubiquitous throughout the Bishop Tuff. Even less abundant, but present throughout the stratigraphy are ilmenite, zircon and apatite. Allanite is present in only earlier-erupted Bishop Tuff, while in contrast clinopyroxene and orthopyroxene are only present in later-erupted Bishop Tuff: Ig2E and all northern-erupted units (Hildreth & Wilson, 2007).

Studies of volatiles in quartz-hosted melt inclusions show a general decrease in water and increase in CO₂ contents and inferred melt inclusion entrapment pressures from earlier, eastern eruptives into later, northern units (Dunbar & Hervig, 1992a; Wallace et al., 1999; Anderson et al., 2000; Roberge et al., 2013). The earlier eastern deposited ignimbrite units, together with their coeval fall deposits, are consequently inferred to represent the upper part of the magma body (depths as shallow as 5-6 km), whereas the later, northern-deposited ignimbrite packages are thought to represent magma erupted from deeper (up to 10-11 km) in the magma chamber (Wallace et al., 1999).

Prior to eruption of the Bishop Tuff the rhyolitic Glass Mountain system was in operation from ~2.2 Ma (and possibly earlier: Chamberlain et al., 2014a) to ~860 ka producing more than 60 effusive and explosive eruptions (Metz & Mahood, 1985, 1991;
Halliday et al., 1989; Metz & Bailey, 1993; Hildreth, 2004; Chamberlain et al., 2014a; Simon et al., 2014). These eruptions built an edifice of domes and associated pyroclastic fans which were partly truncated by the Bishop caldera collapse, but the remaining part of which served to provide a topographic high which divided the northerly and easterly-directed Bishop pyroclastic flows. Recent eruption age estimates [by U-Pb determinations on zircon (Chamberlain et al., 2014a) and ⁴⁰Ar/³⁹Ar age experiments (Simon et al., 2014)] both indicate that eruption of the youngest Glass Mountain dome occurred ~95 kyr prior to the climactic Bishop Tuff eruption.

**SAMPLING AND ANALYTICAL TECHNIQUES**

Sixty samples of the Bishop Tuff were analysed (Table 1), selected to represent all ignimbrite packages, plus fall units 1, 2, 7 and 9 (Fig. 2). Pumice clasts were chosen to span the range of ‘normal’ crystal-poor (xp) to crystal-rich (xr) clasts, as well as the ‘variant’ pumice types, including those labelled as chocolate (dark), swirly, and dacitic, as defined by Hildreth & Wilson (2007). For a full description of pumice textures in the Bishop Tuff see Hildreth & Wilson (2007). Where possible, single pumice clasts were used; however, if this was not feasible, multiple lapilli-sized pumice clasts were collected. All clasts were cleaned of any adhering matrix prior to sample preparation. Full details of all samples collected and analyses undertaken are in Table 1.

Pumice clasts were rough crushed using a Boyd crusher, and sieved to various size fractions. Crystals were hand-picked from the 1-2 mm (feldspars), 0.5-1 mm (biotite, glass) or 250-500 µm (pyroxenes) sieve fractions, mounted into epoxy discs, polished to expose the crystal cores and carbon coated for imaging and analysis. Although pyroxenes, both ortho- and clino-, have been reported from the later-erupted northern material, and some Sherwin subunit (latest Ig1Eb) and later eastern (Ig2Eb,c) pumices (Hildreth, 1977, 1979; Hildreth &
Wilson, 2007), pyroxenes were only recovered from the Ig2NWa, Ig2NWb, Ig2Nb and Ig2Nc samples in this study, with no appropriately sized (≥250 μm) pyroxenes present in any other samples. Major element analyses were obtained on a JEOL JXA 8230 electron probe microanalyser (EPMA) at Victoria University of Wellington (VUW) using wavelength-dispersive spectrometry. Prior to analysis, back-scattered electron (BSE) images were taken of all crystal phases to identify zoning patterns and locate the analytical spots. Precision of standard analyses of major elements (>5 wt.% concentration) is nearly always within 5 relative % (2 s.d.); uncertainties are slightly higher for minor elements (for details see Electronic Appendix 1). For biotite, non-stoichiometric analyses with totals of <95 wt.% and anomalous K$_2$O concentrations were set aside (as per Hildreth, 1977). Similarly, glass analyses with totals of <93 wt.% were set aside; values for the remaining analyses were then normalised to 100 %.

Trace element analyses were measured *in situ* using a New Wave deep UV laser (193 nm solid state) coupled to an Agilent 7500cs inductively coupled plasma mass spectrometer (ICPMS) at Victoria University of Wellington (VUW). The laser ablation (LA)-ICPMS data were internally normalized to $^{29}$Si or $^{43}$Ca from EPMA analyses. Abundances of single trace elements were calculated relative to a bracketing standard (NIST 612 or BCR-2G) which was analysed throughout the run under identical conditions. Precision and accuracies varied depending on the analytical conditions but generally have <5% (2 s.d.) uncertainties on elements with abundances of >20 ppm (slightly higher for less abundant elements: for details see Electronic Appendix 1).

**RESULTS**

**Sanidine**
Sanidine is one of the two most abundant crystal species in the Bishop Tuff, making up ~40% of the crystal population (Hildreth, 1977, 1979; Hildreth & Wilson, 2007). Sanidine crystals from eastern units are dominantly unzoned under BSE (Fig. 3) reflecting the lack of compositional variation measured by EPMA analyses (Fig. 4). Of 28 samples from all eastern ignimbrite and most fall (F1-F8) units, only two (BP032 and BP244) contained sanidine grains with bright rims in BSE imagery (10 and 25% of the grains imaged, respectively: Table 1). The nine samples from F9, Ig2NWa and Ig2Na have 14 to 45% of their sanidine with bright rims, while the eleven late northern (Ig2NWb, Ig2Nb, and Ig2Nc) samples have 45% to 100% of bright rimmed sanidine (Figs. 2 and 3; Table 1). Swirly or dacitic pumices additionally sometimes include anomalous sanidine grains which are commonly fractured, cloudy and completely bright under BSE, and are thus interpreted as xenocrystic (Fig. 3). These grains are not included in the percent bright rim figures listed in Table 1.

Major element \((n = 1403)\) and trace element \((n = 834)\) analyses of sanidine crystals show limited ranges in composition for most elements, with most variations being within the 2 s.d. analytical uncertainty (Table 2; Fig. 4a). 97% of sanidine compositions are \(\text{Or}_{60} \) to \(\text{Or}_{68} \) with variations outside this range found only in xenocrystic grains \(\text{Or}_{>68}\) or rare examples of other overgrowth rims \(\text{Or}_{<60}\), enriched in \(\text{Na}_2\text{O}\), found only in samples from northern vented ignimbrite packages from the north side of the caldera.

In terms of major elements (i.e., those measured using EPMA), sanidine from eastern deposited units (both fall deposits and ignimbrite) has only low-Ba concentrations relative to northern deposited units even if bright rims are visible from BSE images and irrespective of whether cores or rims are analysed. Although commonly enriched overall in Ba (and also Ca) when compared with sanidine in eastern-deposited units, sanidine from the northern vented ignimbrite packages is much more variable in its zoning patterns. These patterns range from low-Ba cores and rims (43% of sanidine), to low-Ba cores and higher Ba rims (47%), to
moderate Ba cores, with higher Ba rims (10%). Sanidine cores in both eastern and northern units are always depleted in Ba relative to their rims. The differences, however, between cores and rims in eastern sanidines are always <3500 ppm Ba, whereas in northern sanidines the core-rim differences are larger, up to 11000 ppm Ba. When all the sanidine major element data are considered as a whole, two modes of composition can be identified using Ba contents (as in Gualda et al., 2012a; Fig. 4b), although it is apparent that these lower-Ba and higher-Ba modes are represented by core and rim analyses in northern sanidines, respectively, often within the same crystal (Fig. 3). Ig2Na sanidine shows the most evolved compositions of samples from northern ignimbrite units, with major element compositions which overlap most closely with sanidine from eastern ignimbrite units.

In terms of trace elements, increases in Ba, Sr, Ti, and the light Rare Earth Elements (LREE) are the most significant, especially in the bright rim overgrowths that are common in sanidine from the later-erupted northern units. Sanidine zones with higher-Ba (dominantly represented by rims in northern vented units) commonly also show enrichments in Mg, Sr and LREE and depletions in Rb (Table 2; Fig.4) relative to the lower-Ba group (typified by cores and rims in eastern vented samples, and cores in northern vented units). Within the lower-Ba grains from eastern ignimbrite units there are still slight rimwards increases in Ba, only resolvable by LA-ICPMS analysis, to maximum concentrations of ~800 ppm, even in grains where BSE-visible bright rims are absent and EPMA could not resolve any changes in Ba concentration. Sanidine from swirly and dacite pumices has trace element characteristics that are indistinguishable from each other, as well as from sanidines in ‘normal’ pumice samples (Fig. 4f). The xenocrystic sanidine grains (identified by BSE imagery), however, have much higher Ba, Sr and Ca concentrations.

**Plagioclase**
Plagioclase is the third most abundant mineral species in the Bishop Tuff, making up ~10% of the crystal population (Hildreth, 1977, 1979; Hildreth & Wilson, 2007). Two groups of plagioclase have been identified using BSE imagery and confirmed with EPMA data. The first group is completely unzoned, whereas the second group shows faint oscillatory zoning (Fig. 3). These oscillatory-zoned crystals are generally only found in the late, northern vented samples, and there form 17 to 100% of the total plagioclase population per pumice sample.

Major- ($n = 251$) and trace-element ($n = 310$) analyses were carried out on Bishop plagioclase (Tables 1 and 3). In tune with observations from BSE imagery, two plagioclase populations are identified based on CaO content, similar to those reported in Gualda et al. (2012b), yet the changes in CaO content recorded by us are continuous, rather than bimodal.

A higher-Ca group of crystals (CaO >4.4 wt.%; >An$_{22}$) is dominated (82%) by the oscillatory-zoned plagioclase, whereas a lower-Ca group (CaO <4.4 wt.%; <An$_{22}$) are dominantly (>65%) unzoned in BSE imagery. Unlike in the sanidine, there are no systematic differences between cores and rims of plagioclase, with only ~5% of crystals showing an overgrowth of a lower- or higher-CaO rim over a higher/lower-CaO core (not shown). Similar to the sanidine, northern-sourced plagioclase represents almost the entire range in CaO content, with only a single eastern-sourced sample (BP198), which has visual evidence for magma mixing in hand sample, reaching the higher-Ca compositions common in northern samples. The lower-Ca plagioclase, present throughout all Bishop Tuff samples, has no identifiable differences in its composition between eastern vented (earlier-erupted) versus northern vented (later-erupted) samples. The higher-Ca plagioclase has corresponding enrichments in Sr, Ba, Ti (Fig. 5) as well as Mg, Eu and LREE when compared with the lower-Ca plagioclase. The lower-Ca plagioclase shows weak positive correlations between Sr and Ba, or LREE, which do not vary based on either stratigraphic position or core-to-rim variations, in contrast to the changes observed in sanidine.
Biotite

Biotite is the most common ferromagnesian phase in the Bishop Tuff, increasing in abundance from <0.05 vol.% in early fall units to ~0.5 vol.% in late northern ignimbrite (Hildreth, 1979). Biotite was extracted from 25 pumices from early fall units to late-erupted ignimbrite (Table 1). Although commonly found as golden brown weathered crystals, every effort to select the freshest looking black biotite was made. Where present, biotite occurs as euhedral crystals which display no evidence for dissolution. When imaged under BSE, no zoning was evident in any crystals. 236 usable major element compositions were determined (where usable here refers to totals >95 wt.%) from fresh biotite crystals. Biotite from early-erupted units (Ig1E, fall units 1-8) systematically produced low totals (Table 4) due to strong interlayer deficiencies in K$_2$O, and so could not be used for accurate comparisons. This alteration has been attributed to syn-eruptive destabilisation of biotite in the upper, more water-rich parts of the Bishop magma chamber (Hildreth, 1977). Although limited by the varying degrees of weathering which they have undergone, later-erupted biotite crystals are enriched in TiO$_2$ and extend to lower K$_2$O values than earlier-erupted biotite. No other differences in major element chemistry can be discerned from the EPMA data (Fig. 6).

Variable degrees of biotite alteration are noted in all samples (similar to Hildreth, 1977, 1979) therefore further investigation using trace element analyses was not undertaken, and the value of biotite in investigating or constraining the nature of the pre-eruptive magma body may be limited.

Orthopyroxene

All orthopyroxene present is euhedral, and in BSE imagery shows no evidence for any late-stage dissolution. Using high contrast BSE imagery, some zoning is discernible, with Mg#
values obtained from subsequent EPMA analyses varying from 48.2 to 56.3 (Fig. 7).

Orthopyroxene displays a range in zoning textures which define four groups: 1) completely unzoned; 2) simple zoned with a BSE-brighter (Fe-richer) rim; 3) simple zoned with a BSE-darker (Mg-richer) rim; and 4) complexly zoned crystals (Fig. 3). The zoning style does not correlate systematically to any particular sample or eruptive unit (Electronic Appendix 2).

Major element analyses \((n = 285)\) of orthopyroxenes show slight variations in SiO\(_2\), MgO, FeO and CaO (Table 5; Fig. 7). Trace element analyses \((n = 269)\) show that Mn and Zn are positively correlated (Fig. 7c), although no other trace elements show systematic variation (Fig. 7e). Interestingly, although slight zoning is evident under BSE imagery, there are no systematic major or trace element differences between cores and rims, even when only cores and rims from one zoning type is considered. Similarly, there are no systematic relationships in orthopyroxene composition with the crystallinity of the host pumice, its location in the eruptive stratigraphy or between different pumice textural groups (e.g. swirly vs. ‘normal’; Fig. 7).

**Clinopyroxene**

Clinopyroxene occurs in similar proportions as orthopyroxene (Hildreth, 1977), and is only found in samples which also contain orthopyroxene. Similar to orthopyroxene, clinopyroxene is euhedral. Clinopyroxene BSE images commonly reveal more well-developed zoning patterns than in orthopyroxene, and again can be divided into 4 groups: unzoned crystals, those with a BSE-brighter rim, those with a BSE-darker rim and those with more complex zoning patterns (Fig. 3).

Major element analyses \((n = 215)\) and trace element analyses \((n = 143)\) of clinopyroxene were obtained. Clinopyroxene varies between 41.7% Wo and 44.2% Wo and major-element concentrations are relatively uniform with only slight variations in SiO\(_2\), CaO,
MgO and FeO (Table 5; Fig. 7b). Trace elements are more variable than in orthopyroxene: Zn correlates negatively against Eu, Zr, Ni and Co, but correlates positively with Mn and Y, with no systematic variation in other trace elements. However, these observed variations do not correlate with core versus rims, pumice type (swirly vs. normal), crystal content, or by north versus northwestern ignimbrite packages (Fig. 7) where, similar to orthopyroxene, clinopyroxene cores and rims show the same ranges in trace element concentrations. Zoning type is also not a control on the observed trace element variations.

**Glass host material**

Fifty-six samples of separated glass fractions from the Bishop Tuff were analysed (Table 1). The data set comprises both major- ($n = 1571$) and trace-element ($n = 697$) analyses. Elements with concentrations close to their detection limits using EPMA were also analysed using LA-ICPMS techniques to improve precision and accuracy.

**Glass chemistry from ‘normal’ early- and intermediate-erupted pumices**

Pumices from F1-F8 and the eastern ignimbrite units (representative of earliest- to intermediate-erupted magma) with a range of crystallinity were analysed (see Table 1). Variations in major element glass chemistry are very restricted in eastern samples, all having SiO$_2$ contents (calculated volatile free) of 75.9-78.4 wt.% ($\pm 1.2$ wt.%, 2 s.d. for SiO$_2$ analyses by EPMA; Table 6). Slight variations in K$_2$O, Al$_2$O$_3$, CaO and Na$_2$O are observed, but these are challenging to resolve beyond analytical uncertainty with EPMA techniques (Fig. 8). Trace element analyses of eastern glass reveal much more distinctive chemical variations and highlight systematic trends. The most evolved compositions, defined by those with the lowest Sr, Ba, LREE, Ca, Ti, Zr and Mg, and the highest incompatible element (Rb, Nb, Cs) contents (Hildreth, 1977, 1979) are found in the very earliest fall material (F1, F2) and some Ig1Ea
samples (Table 6; Fig. 8c). The measured variability within a single pumice increases with progression through the eruption, with some eastern samples displaying a twofold variation in Ba and Sr concentrations from different glass shards yielded by crushing of a single clast (Fig. 8c, d, f). By comparing the compositional data of eastern glass samples with data from glass matrices in pumices from the Ig1NW, Ig1SW and Ig2SW ignimbrite units, it is clear that samples from these units all share trace element characteristics similar to the ‘normal’ eastern type samples (Fig. 8), which includes some limited overlap with the ‘normal’ northern type samples.

Glass chemistry from ‘normal’ later-erupted pumices

Glass matrix samples from pumices in northern ignimbrite lobes (representative of later-erupted material) and unit F9 have more variable compositions than those from the earlier-erupted eastern samples, although SiO$_2$ values (75.4 to 78.3 wt.%, volatile free) cover a similar range. Glasses from northern samples are typically more enriched in Ba, Sr, LREE, Ti and Zr when compared with eastern glass, and similarly are more depleted in Mn, Rb, Nb, Cs, Th and U (Table 6; Fig. 8). Glass samples from Ig2Na and F9 have the most evolved compositions of all later-erupted material, apparently defining an overlap between earlier, more-evolved and later, less-evolved glass compositions (Fig. 8e). Single-pumice variations in glass compositions are also much greater in later erupted pumices than in earlier units, with Ba varying by up to ~100 ppm in glass within a single clast (Fig. 8c, d, f).

Glass chemistry from variant (swirly, dark) pumices

Of the clasts identified as ‘swirly’ or ‘dark’ (Table 1) only five showed any distinct differences from the suite of Bishop Tuff ‘normal’ pumice glasses. Samples BP032, BP197, BP198, BP230 and BP232 all show enrichments in Ca, Fe, Ti, Ba, V, Mn, Sr and Mg.
compared with ‘normal’ Bishop glass (eastern and northern: Fig. 9a, b). There are, however, no variations in any REE, or Rb, Nb and Cs. Compositional variations within single clasts are very large within the swirly pumice suite, greatly exceeding those observed in any of the normal pumices.

Glass chemistry from dacitic juvenile clasts
Three less-evolved samples from Ig2NWb were examined, where black (or red, where oxidised) poorly vesicular material of dacitic composition is comingled with white rhyolite, (the dacites of Hildreth & Wilson, 2007). All three samples have dacite glass compositions which are distinctly different from all others measured in the Bishop Tuff. Instead of a continuum in glass chemistry, as evidenced from variable degrees of mixing in swirly samples, distinct bimodality is observed. The white rhyolite glass component of the dacitic clasts overlaps with the glass compositions of non-mixed pumices in the northern-vented Bishop ignimbrite. In contrast, the darker dacite (glass SiO$_2$ of 64.6 – 69.9 wt.%, volatile free) is significantly enriched (by up to 4 to 5 times: Table 6) in Ba, Sr, Mg, Ti, V, Ca, Zr, Eu and Ni, with accompanying depletions in Rb when compared with rhyolite glass (see Table 6; Fig. 9). This bimodality in glass compositions within single clasts shows that very little chemical mixing occurred between the rhyolite and dacite melts prior to or during eruption.

INSIGHTS INTO THE BISHOP MAGMA BODY
We use our major and trace element analyses of glass and phenocrysts from samples spanning the Bishop Tuff sequence to address four key issues about the Bishop magma system. In conjunction with our new data we utilise previously published data sets of major and trace element concentrations in melt inclusions (Dunbar & Hervig, 1992a; Wallace et al., 1999; Anderson et al., 2000; Roberge et al., 2013), zircon textures and trace element concentrations
(Reid et al., 2011; Chamberlain et al., 2014a) and quartz and sanidine zoning patterns (Anderson et al., 2000; Peppard et al., 2001; Wark et al., 2007; Chamberlain et al., 2014b).

The issues we address are considered under headings as follows.

1). **Origins of Bishop crystals.** Did all the crystals grow in the melt/glass in which they are now found, as proposed by Hildreth (1977, 1979)? If not, then which crystals (or parts thereof) were sourced from what magmas?

2). **Compositional variations in the Bishop magma chamber.** What was the nature of the zonation within the pre-eruptive Bishop magma chamber, and what role did late-stage intrusion of the contrasting magma that gave rise to the CL-bright (zircon, quartz: Peppard et al., 2001; Wark et al., 2007; Chamberlain et al., 2014a) and BSE-bright (sanidine: Chamberlain et al., 2014b) rims have in establishing this zonation?

3). **Sources of subordinate ignimbrite packages.** Where in the magma body did the volumetrically subordinate or poorly preserved Ig1SW, Ig1NW and Ig2SW packages (Fig. 2; Wilson & Hildreth, 1997) originate? How do their compositional data relate to zoning within the Bishop magma chamber?

4). **Geometry of the Bishop magma chamber.** Is there geochemical evidence for a horizontal or stepped roof of the Bishop magma chamber, and how does our new data contribute to a revised magma chamber model?

In the sections below, we discuss the implications of our new data. In a subsequent section we address the current controversies about the Bishop magma system, including a proposal that the Bishop Tuff eruption was the product of two spatially distinct magma chambers.

**Origins of Bishop crystals**
We couple our new crystal and glass data with published information to consider the origins of the common Bishop Tuff crystal phases. Originally the hypothesis was advanced that all the crystals were phenocrysts that grew within the melt in which they were immersed on eruption (Hildreth, 1979) with no evidence of what would now be termed antecrystic major phases (e.g. Bacon & Lowenstern, 2005; Charlier et al., 2005, 2008; Miller et al., 2007; Allan et al., 2013). However, since Hildreth (1979) it has been recognised that many crystals in later-erupted units show a two-step history, precluding interpretation of all crystals as phenocrysts. The late-stage mixing event with the ‘bright-rim’ magma affected the deeper parts of the Bishop chamber, resulting in overgrowth of bright (under BSE or CL) crystal rims, richer in Ba and Sr in sanidine, richer in Ti in quartz, and poorer in U, Th and REE in zircon (Anderson et al., 2000; Peppard et al., 2001; Simon & Reid, 2005; Gualda, 2007; Hildreth & Wilson, 2007; Wark et al., 2007; Gualda et al., 2012b; Roberge et al., 2013; Chamberlain et al., 2014a,b).

Sanidine

Two-feldspar thermometry on plagioclase inclusions within sanidine cores, or sanidine inclusions within plagioclase cores, undertaken on the same samples used in this study, yields a range in model temperatures of ~80 °C (740 °C to 820 °C) from early eastern units to late northern units (Chamberlain et al., 2014b), similar to ranges in Fe-Ti oxide and O-isotopic model temperatures (see below and Bindeman & Valley, 2002; Hildreth & Wilson, 2007). When the previously published model temperatures are compared with our new trace element measurements in the host sanidine cores, those with higher temperatures also tend to have higher Ba, Sr and LREE concentrations (Fig. 10a). Similarly, the cores of all sanidine crystals from samples of later, northern units are enriched in Ba, Sr, LREE and Ti (Fig. 10c), highlighting their less-evolved nature when compared with sanidine cores from samples of
earlier eastern units. These variations and continuity of the data show that the sanidine crystals have been resident in the normal Bishop magma for their entire growth, consistent with melt inclusion compositions in quartz: when inclusions in CL-bright rims are excluded, there is a gradual change from more-evolved compositions in the earlier-erupted units, to less-evolved in the later-erupted samples (Roberge et al., 2013). It is important to note, however, that the cores of sanidine crystals grew in the less-evolved northern melt prior to mixing with the late-stage bright-rim magma, and thus could be considered antecrystic to the final erupted Bishop magma compositions - where only the rims are in equilibrium with the melt in which they were erupted (cf. Hildreth, 1979; Evans et al., 2015).

Plagioclase

Although there are no systematic differences in core compositions of lower-Ca plagioclase crystals from eastern and northern units of the Bishop Tuff (Fig. 10b), their euhedral appearance and the occurrence of plagioclase of similar composition as inclusions within sanidine indicates that the lower-Ca plagioclase crystals are phenocrysts. In comparison, the enrichments in Ti, Sr, Ba in higher-Ca plagioclase might be taken to suggest that this mode of plagioclase may not be phenocrystic, and may have grown either during or after late-stage intrusion of the ‘bright rim’ magma (e.g. Chamberlain et al., 2014b). To investigate this possibility we modelled the melt compositions from which the higher-Ca plagioclase crystallised, using the partition coefficients of Brophy et al. (2011). By comparing the modelled compositions to those measured in glasses from the rare dacite samples (as a proxy for the ‘bright-rim’ magma, see later discussion), we find that the modelled melt compositions cannot match the high Sr content of the dacite glasses (Fig. 11a). Furthermore, the plagioclase inclusions within the cores of non-BSE-bright sanidine crystals include some of higher-Ca composition. These pieces of evidence imply that the higher-Ca plagioclase
crystals, despite their slight compositional differences to the lower-Ca plagioclase, are not simply related to interactions with the ‘bright-rim’ magma, and are themselves also true phenocrysts. We infer that the overall spectrum of plagioclase compositions and generally (but not exactly) matching textures reflect subtle, second-order variations in volatile contents and melt compositions within the Bishop magma body.

Pyroxenes

Pyroxenes in the Bishop Tuff are suggested by some authors to be late-stage entrants into the magma chamber and not in equilibrium with the other mineral phases (Evans & Bachmann, 2013). Equilibrium between the Bishop pyroxenes (and between them and melt) is, however, suggested by their euhedral appearance (Fig. 3), by δ^{18}O values consistent with equilibrium between both pyroxenes and quartz, magnetite, zircon and titanomagnetite (Bindeman & Valley, 2002), and by the compositions of all crystal inclusions within pyroxenes being the same as those of glass-coated crystals (Hildreth, 1977; Hildreth & Wilson, 2007). Here we use our measured trace element concentrations in crystal cores from both ortho- and clinopyroxene [using the partition coefficients of Brophy et al. (2011) and Olin & Wolff (2010), respectively] to model melt compositions from which the crystals grew (Fig. 11).

Clinopyroxene modelled-melt compositions overlap with the Bishop Tuff field, and are distinct from the fields of the dacite compositions (used as proxies for the ‘bright-rim’ magma; Fig. 11b), suggesting that the clinopyroxene is phenocrystic. In contrast, the orthopyroxene modelled-melt compositions, while partly overlapping the ‘normal’ Bishop field, also overlap with the measured dacitic glass (Fig. 11c) suggesting that at least some orthopyroxene is phenocrystic to the ‘bright-rim’ magma and was hence introduced into the pre-existing Bishop magma by magma mixing. Complexities in unravelling melt compositions from crystals arise from the assumption that the dacites represent the particular
melt from which the orthopyroxene crystallised. This may not in fact be the case: the ‘bright-rim’ magma may itself not be definable by one single melt composition, as suggested by the compositional heterogeneities in glasses from dacitic and swirly clasts. Although the two pyroxenes in the Bishop Tuff may have different origins, they are in apparent oxygen isotopic equilibrium (Bindeman & Valley, 2002), and lack evidence for wide-scale dissolution. We attribute this apparent equilibrium to the broadly similar pressures and temperatures of the two melts, and the lack of major compositional and thermal contrasts between the resident Bishop magma and the incoming ‘bright-rim’ magma.

Experimental studies yield contrasting perspectives on the origins of Bishop pyroxenes. Naney (1983) showed that at 200 MPa, high temperatures (~820 – 840 °C) and a water content of ~4.1 wt.% (these pressure and water content values being similar to those inferred from melt inclusions within late Bishop Tuff quartz: Wallace et al., 1999), both orthopyroxene and clinopyroxene would be stable. Conversely, at 200 MPa but with higher water contents and lower temperatures appropriate for modelling the early and middle Bishop Tuff (5.3 wt.% H2O: Wallace et al., 1999; 711–758 °C: Chamberlain et al., 2014b), only clinopyroxene occurred. However melt inclusion entrapment pressures for the early and middle Bishop material are <200 MPa (Wallace et al., 1999), and the lower overall pressures for these materials could explain the absence of clinopyroxene. Scaillet & Hildreth (2001) reported that clinopyroxene (but not orthopyroxene) was stable at 200 MPa using starting composition of Ig2NWb rhyolite, and orthopyroxene (but not clinopyroxene) was stable in a dark variant pumice (67 wt.% SiO2) from Ig1Eb in experiments run at similar P,T conditions to the rhyolites. Note, however, that Scaillet & Hildreth (2001) were unable to crystallise plagioclase or quartz at temperatures >710 °C in their experiments (cf. Naney, 1983).

In summary, our modelling of melt compositions from clino- and orthopyroxene cores is consistent or compatible with the experimental results to suggest that clinopyroxene was a
phenocryst phase in the deeper parts of the Bishop magma body. In contrast, we consider that orthopyroxene was stable in the ‘bright-rim’ magma, and only entered the Bishop magma chamber when mixing commenced. We would thus interpret the small and non-unidirectional zonation patterns seen in the ortho- and clinopyroxene crystals (Fig. 7) to reflect continued crystal growth in subtly contrasting domains of melt composition created during the mixing process.

Summary of crystal origins

Chamberlain et al. (2014a) suggested that zircons composed of 80-100% CL-bright material in pumices from late-erupted units of the Bishop Tuff grew largely or wholly within the ‘bright-rim’ magma (Fig. 12). In contrast, no sanidine or quartz crystals appear to have been phenocrysts resident in the ‘bright-rim’ melt with no purely BSE-bright sanidine found (apart from those with xenocrystic origins, identified by their anomalous appearance and compositions (Fig. 3a) and no purely CL-bright quartz observed (Chamberlain et al., 2014b). This lack is inferred to reflect the slightly higher temperature of the ‘bright rim’ melt (Wark et al., 2007; Chamberlain et al., 2014a) and its less-evolved composition with lower H₂O and higher CO₂, which resulted in sanidine and quartz not being stable phases (cf. Naney, 1983). This inference is in turn consistent with the partial dissolution of cores observed in quartz and sanidine with bright overgrowths (Peppard et al., 2001; Wark et al., 2007; Gualda et al., 2012b; Chamberlain et al., 2014b). After interaction between the ‘bright-rim’ and pre-existing Bishop magmas, some cooling was required to allow resumption of quartz and sanidine crystallisation in the relevant part of the magma body with growth of bright rims from the mixed magma.

Our results and those of other workers cited above therefore suggest that sanidine, plagioclase and quartz are true phenocrysts to the original, zoned Bishop magma body that
was established prior to mixing with the ‘bright rim’ magma. Other accompanying phases in
the early to middle-erupted parts of the Bishop Tuff (allanite, Fe-Ti oxides: Hildreth, 1979;
zircons lacking bright rims: Chamberlain et al., 2014a) would similarly be inferred to be
phenocrysts. This conclusion (matching that of Hildreth, 1979) is consistent with the
interpretation that the Bishop magma body was assembled by piecemeal release of essentially
aphyric (and compositionally slightly heterogeneous) melt batches from a common mush
zone, which then rose to their respective levels of neutral buoyancy based on the slight
compositional differences and underwent some crystallisation (Hildreth & Wilson, 2007).
The clinopyroxene is phenocrystic to the deeper, hotter magma, but available evidence cannot
discriminate whether this phase grew prior to introduction of the ‘bright rim’ magma, or was
in part a product of mixing between the lower Bishop and ‘bright-rim’ magmas, or both (Fig.
12). Orthopyroxene and dominantly CL-bright zircon, in contrast, are inferred to be native
only to the incoming ‘bright-rim’ magma (Fig. 11).

**Compositional variations in the Bishop magma chamber**

*Vertical variations*

As discussed in the previous section, use of two-feldspar thermometry (Chamberlain et al.,
2014b) and trace element analyses within sanidine cores show a systematic change from
more-evolved compositions, lower temperatures in the earlier eastern erupted units to less-
evolved, compositions and higher temperatures in the later, northern erupted units of the
Bishop Tuff (Fig. 10) prior to mixing with the ‘bright-rim’ magma. These variations in and
continuity of the data show that the Bishop Tuff magma chamber was thermally and
compositionally zoned before any late-stage mixing with the ‘bright-rim’ magma (Fig. 12)

To constrain the zonation within the magma chamber after mixing with the ‘bright-
rim’ magma and immediately prior to eruption we use glass compositions from all units of
the Bishop Tuff. On the large scale, when considering only pumices which fall into the
‘normal’ xp to xr spectrum of Hildreth & Wilson (2007), a systematic progression is evident
in the glass trace-element data, from the most-evolved Ba, Sr, Ti-poor melt in the earliest fall
units, to less-evolved, Ba, Sr and Ti-enriched melt in the order of: early fall (F1- F6) – early
eastern ignimbrites (Ig1E) and F7 - F8 – late eastern units (Ig2E and Ig2SW) – Sherwin lobe
(geographically confined lobe of latest Ig1Eb) – Ig2Na (and F8/9) – Later northern units
(Ig2Nb/c; Ig2NW) (Table 6; Figs. 8, 13; Electronic Appendix 2). Although the Sherwin lobe
underlies Ig2E, in the geographically constrained segment of the chamber tapped by the
Sherwin lobe the melt was less-evolved than that sourcing Ig2E. The least-evolved glass
compositions (77.8 wt.% SiO$_2$; 0.50 wt.% CaO; 950 ppm Ti; 55 ppm Sr; 340 ppm Ba: Table
6) in normal pumices are found in Ig2Nb/c and Ig2NWa/b samples, with no systematic
difference between these units. However, given the mineralogical evidence for interaction
with the ‘bright-rim’ magma in these units, their glass compositions are interpreted to reflect
a mixture of the less-evolved Bishop and ‘bright-rim’ magmas (Fig. 12). The general
progression in glass compositions is in agreement with those reported from previous analyses
of matrix glass, melt inclusions, and whole rock samples (Hildreth, 1979; Lu et al., 1992;
Wallace et al., 1999; Anderson et al., 2000; Hildreth & Wilson, 2007; Roberge et al., 2013).

On a finer scale, we note that samples from Ig2Na (Figs. 6, 8, 11) share several
characteristics in common with samples from both northern (later) and eastern (earlier)
ignimbrites. Ig2Na glass chemistry overlaps both northern and eastern glass compositions
(Fig. 8e), with melt inclusions from this unit having Rb/Ce and Nb/Ce ratios that span almost
the entire range of matrix glasses from the Bishop Tuff (Roberge et al., 2013). Similarly to
eastern samples, plagioclase compositions are dominantly lower-Ca and pyroxene crystals do
not exceed 250 µm across, while the bright rims on sanidines and zircon crystals are less well
developed when compared with other northern samples (Table 1; Chamberlain et al., 2014a,
This evidence points towards the intermediate nature of Ig2Na and serves to highlight the sequential nature of the mineralogical and geochemical changes from earlier, eastern-erupted, to later, northern-erupted units of the Bishop Tuff. Identification of the more limited role of the ‘bright-rim’ magma in Ig2Na pumices when compared with other northern-sourced ignimbrites, combined with restricted evidence for mixing preserved in melt inclusions from Ig2Ea/b (which varies with geographic location: Roberge et al., 2013), show the diffuse nature of the transition zone between the earlier- and later-erupted materials (Fig. 12). Thus, although evidence for significant mixing is restricted to the northern-erupted units, some indications of mixing are also present in later eastern-erupted units.

**Lateral variations**

Although the Bishop Tuff includes geographically confined ignimbrite packages, some with distinctive pumice suites (e.g. the Sherwin and Watterson subunits of Ig1Eb: Hildreth & Wilson, 2007), lateral variations within the Bishop magma body have not previously been investigated. By collecting samples from multiple locations within single defined units we here test what lateral variations existed (or were preserved during eruption) within the Bishop magma body (Table 1). In most units of the tuff, no identifiable lateral variability can be identified; for example, the Watterson subunit, distinguished from other Ig1Eb samples in the basis of its relatively high abundance of swirly pumices, is indistinguishable based on chemical composition (see Electronic Appendices 1, 2). Samples from the Sherwin subunit of Ig1Eb do, however, show some chemical differences from the rest of Ig1Eb. Glasses from the Sherwin subunit plot in comparable fields to those from Ig2E (Fig. 13), and extend the less-evolved compositions reached by eastern material (though not reaching the high Ba, Sr, LREE levels of northern material). Sample BP040 in the Sherwin subunit displays the common presence of thin CL-bright rims on zircons, in contrast to other Ig1Eb zircons where
such rims are absent (Chamberlain et al., 2014a). This subunit has also been noted to contain some pyroxene-bearing pumice (Hildreth, 1977). However, in all three samples from this unit no pyroxene >250 µm was recovered, similar to Ig2Na samples. The lack of zoned sanidine phenocrysts, or all-bright zircons in Sherwin subunit samples suggests that although some ‘bright-rim’ magma may have reached this sector or level within the magma chamber, it was not volumetrically significant (Fig. 12).

Some lateral variability is also evident in the northern ignimbrite, where packages Ig2NWa and Ig2Na were erupted largely coevally, albeit from different vents (Wilson & Hildreth, 1997). Samples from Ig2NWa are much more typical of the late northern units with prevalent zoning in sanidine, quartz and zircon, high proportions of high-Ca plagioclase, the presence of large (>250 µm) pyroxenes and the less-evolved nature of matrix glasses. We infer that the ‘bright-rim’ magma mixed more freely with the source regions of Ig2NWa than Ig2Na, even though melt inclusion data show broadly similar depths of origin for both of these units (Wallace et al., 1999; Anderson et al., 2000). These observations imply that some lateral heterogeneity was present that is interpreted to reflect uneven degrees of interaction of the ‘bright-rim’ magma within the main Bishop magma body (and variable development of the transition zone: Fig. 12).

Role and nature of the ‘bright-rim’ magma

Late-stage introduction of the ‘bright-rim’ magma affected the lower regions of the Bishop magma body (Hildreth & Wilson, 2007; Wark et al., 2007). Although the ‘bright-rim’ magma played a role in accentuating the vertical zonation in the chamber, it was not the sole or dominant cause for this zonation (cf. Dunbar & Hervig, 1992a; Bindeman & Valley, 2002), as indicated by the gradient in sanidine core compositions and two-feldspar model temperatures (Chamberlain et al., 2014b). Magma erupted from northern vents is thus a mixture of less-
evolved (evidenced by sanidine core compositions) Bishop rhyolite magma and, by comparison, even less-evolved ‘bright-rim’ magma. Caldera-forming eruptions frequently preserve evidence for immediately pre-eruptive introduction of less-evolved magmas which is often cited as an eruptive trigger (e.g. Shane et al., 2005; Wilson et al., 2006; Wark et al., 2007; Wilcock et al., 2013; Goff et al., 2014). However, in the case of the Bishop Tuff, timescale estimates from diffusion studies (Chamberlain et al., 2014b) demonstrate that this mixing with the major mineral phases occurred over a period of up to 500 years prior to eruption and therefore cannot have been the immediate eruption trigger (cf. Wark et al., 2007). Furthermore, age and textural evidence from zircons partially or wholly grown from the ‘bright-rim’ magma suggest that interaction with deep roots of the Bishop system possibly occurred up to 10 kyr before eruption (Chamberlain et al., 2014a).

The materials sampled in the Bishop Tuff that may most closely approach the end-member contributor to the ‘bright rim’ magma are the rare, poorly vesicular dacitic juvenile clasts found in unit Ig2NWb, intermixed with ‘normal’ Bishop rhyolite in hand sample. We infer these dacites represent a very late-stage ingress into the lower parts of the Bishop magma chamber due to the clear bimodality in glass chemistry within single samples (Fig. 9), with no apparent exchange of any elements between the dacite and rhyolite portions of the samples (in blue in Fig. 9). These dacites are the most primitive juvenile material found in the Bishop ignimbrite with the least-evolved (BP211) having 65.65 wt.% SiO$_2$, 4.29 wt.% CaO, 3830 ppm Ti, 360 ppm Sr and 750 ppm Ba (Table 6). The dacites also plot as end-members for the less-evolved (clearly mixed) swirly and dark pumices [such as the sample B443 investigated by Scaillet & Hildreth (2001): 67.26 wt.% SiO$_2$, 3.87 wt.% CaO, 460 ppm Sr and 1350 ppm Ba; Hildreth & Wilson (2007)] (Fig. 9) which are present in variable (but small) proportions throughout the Bishop Tuff. Other dacite, swirly and dark pumices represent a broad compositional continuum (Hildreth & Wilson, 2007; Fig. 9) from the least-evolved
dacite, suggesting that either: 1) varying degrees of mixing with the melts in the main Bishop magma body occurred, or 2) sample variability reflects heterogeneity in the compositions of late intruded magmas and hence that the ‘bright-rim’ magma cannot be defined by a single composition. Given the extended time period over which interactions are inferred to have occurred, both explanations seem plausible.

Wark et al. (2007) specifically hypothesised that a deep intruding CO$_2$-rich mafic magma provided a heat source for dissolution of sanidine and zircon crystals (to supply Ba, Sr and Zr) in an intermediate crystal cumulate (mush) zone between the melt-dominant Bishop magma body and the intruded mafic magma. This melted material then itself was intruded into the overlying magma body, triggered partial dissolution of quartz and feldspar through higher temperatures and then initiated rim overgrowths on these crystals through lowering of the water activity in the melt via CO$_2$ enrichment. This specific hypothesis, however, is at odds with two lines of evidence. First, this process would not explain the increased concentrations of both Ti and Zr in the dacites (see Electronic Appendix 2). Second, the cores of CL-bright rimmed zircons in the Bishop Tuff samples are invariably euhedral, implying that at no stage did the ‘bright-rim’ melts reach temperatures or compositions sufficient to cause zircon dissolution.

The presence of pre-and post-caldera dacites of similar compositions to those erupted with the Bishop Tuff (Bailey, 1989; Hildreth, 2004) imply that the ‘bright-rim’ magma is not unique to the Bishop Tuff, and in fact may represent a more long-lived but sparsely erupted magma type in the Long Valley region. Identification of CL-bright rims on some zircons from dome OD of Glass Mountain (Chamberlain et al., 2014a) suggest also that ancestral ‘bright-rim’ magmas interacted with the Glass Mountain magmatic system. It is therefore plausible that dacite was always present as part of the fractionating sequence of magmas present during rhyolitic activity, but that it could not be erupted through the density trap in
the upper mush zones of the Glass Mountain and Bishop Tuff systems (cf. Hildreth, 2004, his figure 7). Instead it was only able to be erupted at the surface, or mixed with an existing rhyolitic magma chamber, when some other control, possibly tectonic, permitted its passage through the density trap.

Sources of subordinate ignimbrite packages

Three less voluminous (or less well preserved) ignimbrite packages occur: Ig1NW, Ig1SW and Ig2SW (Fig. 2). Given their geographically limited extent these units are often overlooked, but could still yield important insights into the finer stratification of the Bishop magma chamber. By combining our new crystal and glass data with previous field observations, Ig1SW and Ig1NW are linked to a similar magma source as Ig1Eb (Figs. 4, 8; Wilson & Hildreth, 1997; Hildreth & Wilson, 2007). This inference is supported also by the similarity of Fe–Ti-oxide model temperatures and Rb/Sr systematics between these three units (Hildreth, 1979; Christensen & DePaolo, 1993; Hildreth & Wilson, 2007). Ig2SW was originally interpreted to originate from a similar source region to Ig1SW, but from a magma that had slightly higher Fe–Ti-oxide temperatures, zircons with bright overgrowth rims (on 57% of all zircons) and minor pyroxene, indicating similar magmatic conditions to the pumices examined from Ig2E (Hildreth, 1979; Hildreth & Wilson, 2007; Chamberlain et al., 2014a). Our new glass and crystal compositional data confirm this interpretation. The reported presence of minor pyroxene (Hildreth, 1979) and bright rims on zircons (Chamberlain et al., 2014a) raises the possibility that Ig2SW tapped a magma similar to that of the Sherwin subunit, which may thus have extended to around the SW quadrant of the Long Valley caldera.

Geometry of the Bishop magma chamber
Here we collate our new results and previously published material to propose a revised model for the pre-eruptive Bishop magma body. Our sanidine data highlight that this body was zoned in temperature (Chamberlain et al., 2014b) and composition, prior to growth of any crystal phases or any interaction with the ‘bright-rim’ magma, as previously proposed by others (Hildreth, 1977, 1979; Wallace et al., 1999; Anderson et al., 2000; Hildreth & Wilson, 2007). We use the melt inclusion compositional data and entrapment pressure estimates of Wallace et al. (1999), Anderson et al. (2000) and Roberge et al. (2013) to constrain the depths of the magma body. It is established that the earlier-erupted units were largely, but not entirely, sourced from vents along the southern and southeastern sides of the caldera ring fracture (F1-F9 Ig1E; Ig2E; Ig1SW; Ig2SW; Ig1NW) and are volumetrically dominant (roughly two-thirds) when compared with the later-erupted northern units (Hildreth & Wilson, 2007). Our model is similar in many respects to that previously proposed, but incorporates the following key aspects:

1) An asymmetric roof between the southern/eastern and northern vents, following after and scaled from the model of Wallace et al. (1999). Crystal-poor (xp) pumice is rare (<20 %: Hildreth & Wilson, 2007) in northern-erupted units. An important question in constraining the shape of the chamber roof (cf. Fig. 13 of Wallace et al., 1999 versus Fig. 18 of Hildreth & Wilson, 2007), is whether this paucity of xp material represents a pre-eruptive lack of xp, ‘eastern-type’ magma under the northern vents, or if the xp material had originally been present but was drained laterally and erupted from the earlier vents (Hildreth & Wilson, 2007). We thus measured glass compositions from xp pumices from the northern ignimbrite, to determine if any had ‘eastern’-type glass chemistry, or were more evolved than glasses in the co-erupted xr pumices. The data show (Fig. 15) that xp pumices in the northern ignimbrite lobes have similar (or less-evolved) glass (and sanidine) compositions to the xr pumices (similar to those of northern units, Fig. 8), with no clear correlation between crystallinity and
the compositions of matrix glasses or crystal textures (Table 1; Fig. 15). The xp pumices in northern units thus simply reflect a lack of crystallisation or some degree of crystal-liquid separation from the xr magma. These results imply that the ‘eastern-type’ magma was never resident in any significant quantity under the northern vents, with the most-evolved magma being that of the transitional zone, represented by Ig2Na (cf. Hildreth & Wilson, 2007), and are consistent with a stepped upper roof to the magma chamber. Roberge et al. (2013) reported xp pumices from Ig2NWa with quartz-hosted melt inclusions with evolved compositions similar to those in the early erupted materials, but concluded that the host pumices were accidentally incorporated, and not derived from magma underneath the northern rim of the caldera.

2) A transitional zone highlighting the compositional continuum between upper and lower regions of the single magma chamber (cf. Gualda & Ghiorso, 2013a: see later discussion). This transitional region is represented by material forming Ig2Na (and to a lesser degree the Sherwin unit of Ig1Eb), and covers the change from the eastern, more-evolved, dominantly crystal-poor material, with unzoned phenocrysts and no pyroxene, to the northern, less-evolved, dominantly crystal-rich material, with zoned phenocrysts and pyroxenes. The transition zone (Fig. 14) is represented by magma that has sparse, thin CL- or BSE-bright rims on sanidine, quartz and zircon and thus is interpreted to have had shortly pre-eruptive interaction with material that had experienced some mixing with the ‘bright-rim’ magma. The presence of such a transition zone is also consistent with the melt inclusion compositional data of Roberge et al. (2013).

3) A stepped lower margin to the Bishop melt-dominant volume tapped during the eruption (Fig. 14). The physically and compositionally distinctive ‘Adobe-type’ (Hildreth & Wilson, 2007) xr pumices are rare in the eastern units, and therefore it seems improbable that the corresponding less-evolved magma ever underlay the southern and eastern sectors of the
caldera in significant quantities. Crystal-rich pumices investigated from eastern-deposited units (with the exception of those from the Sherwin subunit) all have more-evolved, restricted glass and crystal compositions when compared with northern deposited samples (see Electronic Appendices 1, 2).

4) Diverse intrusions of the compositionally heterogeneous ‘bright-rim’ magma. The less evolved compositions represented by the rare mixed dacitic pumices from Ig2NWb (themselves a possible proxy for the ‘bright-rim’ magma) are inferred to have been fed from depth under the northern roots to the magma body. The ‘bright-rim’ magma itself likely represents multiple injections of compositionally heterogeneous magma over an extended time period, based on the diversity in glass compositions (Fig. 9) and the varied timescales for interaction with the main Bishop magma body (Chamberlain et al., 2014a,b). Variations in composition of the ‘bright-rim’ magma are schematically represented in Fig. 14 by various stalling levels within the underlying mush pile. Any diversity in the range of crystallisation and mush mixing/assimilation would produce slightly different compositions of the resulting magma that eventually reached levels where it mixed in with the overlying Bishop magma body or was captured as ejecta in the eruption.

CONTROVERSIES IN THE BISHOP TUFF

There are two areas of current controversy in and around interpretations of the record preserved in the Bishop Tuff. The first concerns the Fe-Ti oxides, whether or not they are in equilibrium with each other or the host melt and the validity of the model temperature estimates derived from co-existing magnetite-ilmenite pairs. The second concerns the vertical dimensions of the Bishop magma chamber, and whether it was one unitary body, or two separate bodies.
Fe–Ti-oxides – equilibrium or disequilibrium?

A major question regarding the origins of crystals in the Bishop Tuff has been the relationship between the pyroxenes, Fe–Ti-oxides and the surrounding melt (e.g. Hildreth, 1977, 1979; Lindsley et al., 1990, 1991; Frost & Lindsley, 1992; Hildreth & Wilson, 2007; Evans & Bachman, 2013; Ghiorso & Gualda, 2013). As discussed above, we conclude that the pyroxenes are phenocrystic to certain portions of the Bishop magma body. Here we consider the disputes around the validity of the record in Fe-Ti oxides (Ghiorso & Gualda, 2013; Evans & Bachmann, 2013; Evans et al., 2014) and consequent inferences about the thermal structure of the pre-eruptive Bishop magma body.

Analysis of coexisting Fe–Ti-oxide pairs using the Andersen & Lindsley (1988) calibration showed a range in model temperatures from 703 °C in early fall units, to 816 °C in late-erupted ignimbrite samples (Hildreth & Wilson, 2007). Closely similar model temperatures were reported by Ghiorso & Evans (2008) using their modified calibration and these results are reiterated by Evans & Bachmann (2013) and Evans et al. (2014). The general values and ~100 °C range in Fe–Ti-oxide model temperatures from bottom to top of the Bishop Tuff are consistent with those derived from $\delta^{18}$O thermometry utilising the fractionation of $^{18}$O between quartz and magnetite (715 °C – 815 °C; Bindeman & Valley, 2002). Applications of Ti-in-quartz and Ti-in-zircon thermometry also yielded a ~100 °C temperature gradient (Wark et al., 2007; Reid et al., 2011) but the veracity of these model values is challenged by the assumptions associated with the appropriate value for $a\text{TiO}_2$ (see Thomas & Watson, 2012; Wilson et al., 2012) and variations of Ti concentrations between sector zones in zircon (Chamberlain et al., 2014a). Two-feldspar thermometry (Chamberlain et al., 2014b) yields a ~80 °C range in model temperatures from Ig1Eb to Ig2Nc for the cores of crystals, somewhat less that the ~100 °C ranges from other phases, but indicating the
temperature zonation within the magma chamber prior to any changes associated with influx of the ‘bright-rim’ magma.

Use of Rhyolite-MELTS modelling on Bishop Tuff samples has been used to propose that the crystallisation temperature for Bishop Tuff crystals lies within a 10 °C range (Gualda et al., 2012a). The subsequent assertion from Rhyolite-MELTS modelling by Ghiorso & Gualda (2013) that the Bishop Fe–Ti-oxides are not in equilibrium is based on the relationship between calculated model temperatures and modelled $a\text{TiO}_2$, which does not fall within a plausible pressure range as considered by them for the Bishop Tuff. Therefore it was asserted (following Lindsley et al., 1990, 1991; Frost & Lindsley, 1992) that the ilmenite in the Bishop Tuff is very late stage in origin and has not had a chance to equilibrate with magnetite (Ghiorso & Gualda, 2013). Three points suggest strongly, however, that such an argument is not soundly based. First, there is agreement in model temperatures and their ranges between two-feldspar thermometry, $\delta^{18}$O thermometry and Fe–Ti-oxide thermometry. Second, given the demonstrably rapid timescales for equilibration of Fe–Ti-oxides with their host melt (Buddington & Lindsley, 1964; Hammond & Taylor, 1982; Ghiorso & Sack, 1991; Venezky & Rutherford, 1999), the systematic variations between model temperatures and minor element concentrations in both oxide phases (Hildreth, 1979), coupled with their generally euhedral shapes (indicating no widespread dissolution prior to eruption) it would seem most improbable that the magnetite and ilmenite are not in equilibrium (Evans & Bachmann, 2013; Evans et al., 2015). Third, Ghiorso & Gualda (2013) calculate temperature estimates for oxide pairs using the (equilibrium) model of Ghiorso & Evans (2008), yet the resulting derived values of $a\text{TiO}_2$ are assumed to fall outside of their interpreted range for equilibrium pairs and hence to not represent equilibrium behaviour. However, the interpreted range for $a\text{TiO}_2$ is not independently corroborated and therefore is to be a less rigorous test of equilibrium than others available (e.g. Mg-Mn exchange- Bacon & Hirschman, 1988). Our
conclusion is that the Fe–Ti-oxides were in equilibrium with the melt in which they were erupted (see also Evans & Bachmann, 2013; Evans et al., 2015).

**One chamber or two?**

Recent studies have suggested a contrasting model to that in which the Bishop Tuff represents tapping of a unitary zoned magma body (Hildreth, 1979; Wallace et al., 1999; Anderson et al., 2000; Hildreth & Wilson, 2007). Instead it has been proposed by Gualda et al. (2012a), Gualda & Ghiorso (2013a) and Gardner et al. (2014) that the Bishop Tuff is the product of two laterally separated magma bodies, one feeding the early erupted, eastern material, and the other feeding the late-erupted northern material, the latter with higher-Ba sanidine and higher-Ca plagioclase, but with both bodies stored at similar temperatures and pressures.

Three lines of evidence were put forward in support of this two magma chamber hypothesis, which we address here.

(1) A proposed lack of a thermal gradient in the pre-eruptive Bishop magma chamber(s), with all modelled units having temperatures of 740 – 750 °C, according to Rhyolite-MELTS model thermometry and experimental petrology (Gualda et al., 2012a; Gualda & Ghiorso, 2013a; Gardener et al., 2014). However, as discussed above, we (and other workers) find no evidence to suggest that the Fe–Ti oxides were not in equilibrium with the magma that they were erupted with. Both the temperatures modelled from Fe–Ti oxide compositions and those derived from all other mineral-specific methods yield estimates of a ~100 °C gradient in temperature for the immediately pre-eruptive magma body. All mineral-specific and isotopic data serve to highlight the thermally zoned nature of the Bishop magma body.

(2) A proposed narrow range of pressures (175-200 MPa for all samples from the Bishop Tuff) at which the eastern (early) and northern (late) magmas were stored before
eruption, based on Rhyolite-MELTS modelling and Si in glass barometry (Gualda & Ghiorso, 2013a, b). This narrow range in storage pressures is, however, not in agreement with data and inferences made by earlier workers. In particular, studies of quartz-hosted melt inclusions have shown a sequential change in inferred entrapment pressures, from ~160 MPa in the early Bishop Tuff, to ~270 MPa in the late Bishop Tuff, the latter as inferred from melt inclusions trapped in the bright-CL rims (Wallace et al., 1999; Peppard et al., 2001; Roberge et al., 2013). This sequential change in entrapment pressure is mirrored by changes in both melt inclusion and matrix glass compositions, again demonstrated by the matrix glass data presented in this paper. The change from shallower entrapment depths to deeper entrapment depths is not abrupt and in particular the intermediate entrapment pressures of ‘middle-erupted’ eastern material (Ig2Ea) are identified by Wallace et al. (1999). In addition, simple consideration of the volume of the Bishop Tuff divided by the area of the caldera ring fracture yields an average estimate for drawdown in the magma chamber of 2.7 km (Hildreth & Wilson, 2007), equivalent to an average ~60 MPa pressure difference between the earliest and latest-erupted material. Given the inferences we derive in this paper for the stepped floor and roof to the Bishop magma chamber, then this average pressure difference has to represent a minimum value.

(3) An apparent bimodality in crystal compositions [plotted from Hildreth’s (1977) data], is used to suggest crystallisation in separate magma chambers (Gualda & Ghiorso, 2013a). This apparent bimodality, however, simply represents the influence of the ‘bright-rim’ magma on the later erupted Bishop magma, and can be accurately replicated by core and rim analyses from single crystals from single pumices in the northern-erupted units (e.g. Fig. 3). Field evidence also supports a single zoned magma chamber, with the systematic and gradual changes from dominantly crystal-poor pumice in early units, to dominantly crystal-rich pumice in late units (Wilson & Hildreth, 1997; Hildreth & Wilson, 2007), including the
Published evidence thus overwhelmingly implies that the Bishop Tuff eruption systematically tapped a unitary, gradationally zoned magma body. Why Rhyolite-MELTS-derived model-temperature estimates contrast so strongly with other thermometry methods raises questions about its applicability in this case, which could be related to values derived or adopted for $aTiO_2$ and/or the model calibration. Rather than recalibrating MELTs over a wide spectrum of melt compositions, Gualda et al. (2012a) made small changes [comparable in magnitude to the fitting errors in the original MELTs calibration: Ghiorso & Sack (1995)] to the enthalpies of formation for quartz and the potassium end-member of alkali feldspar so that an early Bishop Tuff composition would have quartz, sanidine, and plagioclase on the liquidus. However this approach yields a temperature for early Bishop Tuff magma of ~760°C at 175 MPa and H$_2$O-saturated conditions, higher than the experimentally constrained temperature of <730°C at 200 MPa for the same melt composition (Scaillet & Hildreth, 2001). Gualda et al. (2012a) noted this discrepancy and stated that the offset was likely to be systematic. The absolute accuracy of temperatures calculated with Rhyolite-MELTS should therefore be viewed in this light. Furthermore, Rhyolite-MELTS, as an equilibrium thermodynamic model, is not applicable to samples that represent mixed magmas. Given the evidence for mixing with the ‘bright-rim’ magma (Figs. 4, 8, 9, 12, 13 and 14), Rhyolite-MELTS is unsuited for modelling temperatures and phase assemblages in matrix glass or bulk-rock compositions in the later-erupted Bishop Tuff.

The newly proposed silica-in-glass barometer (Gualda et al., 2013b) also gives similarly restricted ranges in storage pressures, but we consider these results to be overshadowed by dependence of the model on the precision of SiO$_2$ measurements (normally from EPMA). Typical values of 2 s.d. uncertainties vary from ± 0.6 wt.% SiO$_2$ (e.g. Wallace...
et al., 1999; Bachmann et al., 2002; Allan, 2013) to ± 2.0 wt.% SiO$_2$ (Liu et al., 2006; Cooper, 2014). For the former figure, the SiO$_2$-in-glass barometer yields approximate differences in model pressures of ±50 MPa; for the glass data reported in this paper the uncertainty of ±1.2 wt.% SiO$_2$ corresponds to a model pressure uncertainty of ±120 MPa. Pressure differences of less than ~200 MPa are thus in essence unresolvable, given typical uncertainties associated with matrix and melt inclusion glass analyses by EPMA.

Experimental results reported by Gardner et al. (2014) were also used by them to assert that the ‘Late Bishop Tuff magma body’ had to be at a temperature of ≤740 °C, and that higher model temperatures derived from the Fe-Ti oxides were in error. We would suggest that such a conclusion is incorrect for two reasons. First, the bulk composition used in their experiments was that of a whole-rock sample (which is itself a mixed magma) and does not accurately reflect the melt composition of the later-erupted Bishop magma as indicated by melt inclusion compositions (Wallace et al., 1999; Roberge et al., 2013). Second, most of the experiments were carried out under water saturated conditions, despite the melt inclusion data indicating that substantial amounts of CO$_2$ were present in the volatile phase and hence that water activity would have been <1.0. Compositions of late Bishop Tuff melt inclusions are, however, closely matched by a melt composition (AB401) used in the mixed-volatile experiments of Klimm et al. (2008; see also Evans et al., 2015). Their results show that for this composition run under comparable water contents (3-4 wt %) and activities at a pressure of 200 MPa, the three major phases quartz, plagioclase and sanidine are stable together up to temperatures of 775 °C, and the experimental melt compositions under these conditions are a very close match to those of the natural Bishop melt inclusions. Given that the experimental melts do not vary widely with temperature and water content, we infer that at 200-275 MPa, appropriate for the deeper parts of the Bishop magma body shortly before eruption, the match would be as good.
COMPARISON AND CONTRASTS WITH OTHER SUPERVOLCANIC SYSTEMS

The work presented in this paper, combined with recent studies of the Bishop Tuff allows for comparisons and contrasts with other large-scale silicic eruptions (Table 7) to determine if common processes or timescales are present in these super-sized systems.

1). An underlying crystal-rich mush region is inferred to play varying roles in sourcing both melt and crystals into the overlying melt-dominant magma body (see Table 7; e.g. Wilson & Charlier, 2009; Allan et al., 2013; Cooper & Wilson, 2014) and in some cases can demonstrably be shown to be long-lived regions in the crust (Table 7). However, in the Bishop Tuff there are very few zircons with textures or ages appropriate to derivation from the Glass Mountain magma system (Chamberlain et al., 2014a). This observation highlights that either the underlying crystal mush was effectively reset, or that a completely new crystal mush system was established in the ~95 kyr time break between the youngest Glass Mountain and the climactic Bishop Tuff eruptions (Chamberlain et al., 2014a; Simon et al., 2014). The Bishop Tuff samples also show no clear evidence for texturally distinctive crystals that can be interpreted to have originated in a mushy region (cf. Bindeman et al., 2008; Allan et al., 2013).

2). A thermal and compositional gradient was largely established in the Bishop chamber prior to any mixing with the ‘bright-rim’ magma. The preservation of this gradient in systematic crystal and glass compositional variations shows that convection and mixing was not a significant chamber-wide process in the Bishop system, even though it is generally considered to be an inevitable consequence in such large bodies of thermally zoned magma [e.g. Huppert & Turner (1991) and references therein]. The volumetrically dominant upper part of the Bishop magma body thus conforms with the Hildreth (1979) concept of a body that was not convecting on any gross scale, and within which the associated crystal cargo
grew as phenocrysts. The lower part of the Bishop magma body, in contrast, was undergoing mixing and associated convection, with incorporation of various components from the pre-existing Bishop magma and the incoming ‘bright-rim’ magma, and limited amounts of xenocrystic material [e.g. Triassic zircons: Chamberlain et al. (2014a)]. In contrast, magma bodies assembled as rapidly as the Oruanui, with associated high rates of heat input and output, show evidence for wholesale convection in their lack of systematic compositional zonation, and the diversity of crystal and melt-inclusion compositions found in single pumices (Liu et al., 2006; Wilson et al., 2006; Charlier et al., 2008; Allan, 2013; Allan et al., 2013).

3). Zircon U-Pb age data imply that it took ~80 kyr to assemble the >600 km³ melt-dominant Bishop magma body (Chamberlain et al., 2014a). There are no grounds for supposing that accumulation of this body was anything other than steady, based on the general textural uniformity of zircon grains (Chamberlain et al., 2014a), accompanying its proposed piecemeal assembly (Hildreth & Wilson, 2007). In other regions, the central Taupo Volcanic Zone (TVZ) of New Zealand especially, accumulation processes are an order of magnitude or more faster, reflecting the very high magma-related heat fluxes in the region, and commensurately rapid rates of heat loss via geothermal systems (Bibby et al., 1995; Wilson & Charlier, 2009: Table 7). At one extreme, the Oruanui eruption, broadly comparable in size to the Bishop Tuff, had its melt-dominant body assembled in <3000 years, and the underlying magmatic roots were wholly reorganised over a <~5000 year period into a new, highly productive system (Wilson & Charlier, 2009; Allan et al., 2013; Barker et al., 2014).

4). There is a growing contrast between long-established precepts about the prolonged lengths of time required to develop large silicic magma bodies for eruption (e.g. Smith, 1979; Shaw, 1985; Trial & Spera, 1990; Reid, 2008) and data arising from studies involving dating
of eruptive components, especially U-Th disequilibrium and U-Pb dating of accessory phases (e.g. Vazquez & Reid, 2002, 2004; Simon & Reid, 2005; Bindeman et al., 2006, 2008; Wilson & Charlier, 2009; Chamberlain et al., 2014a). In part, these contrasts arise because different processes involved in magma generation and accumulation are being measured. The timescales required to physically accumulate an eruptible body of magma can, in cases like Taupo, be demonstrated to be 1-2 orders of magnitude shorter than the timescales recorded in ages of crystals that reflect the chemical processes that gave rise to those crystals and their associated magma composition (Sutton et al., 2000; Wilson & Charlier, 2009; Allan et al., 2013; Barker et al., 2014). The distinction is recognised at Taupo by the exceptionally frequent snapshots afforded of the magma system, contrasts in the zircon model-age spectra between closely spaced eruptions from the same magma system, and the systematic core-rim relationships in common mineral phases. In the Bishop Tuff, such a distinction is not feasible. The Bishop zircon age spectrum is almost entirely contained within the time gap since the last Glass Mountain eruption, and non-rimmed zircons from the early-erupted Bishop samples show no age or textural characteristics that would allow the contrast between growth in the mush versus growth in the melt-dominant body to be recognised. Overall, accumulation of the Bishop magma body is inferred by us to have occupied the ~80 kyr period indicated from the zircon age spectra, with input from the ‘bright-rim’ magma starting up to 10 kyr before eruption, and escalating in intensity ~500 yr before eruption (Chamberlain et al., 2014a, b). We would suggest that the shorter timescales proposed from Ti diffusion profiling in quartz (Gualda et al., 2012a) reflect the instability (i.e. growth and dissolution) of individual crystals in the later stages of accumulation of the melt-dominant body, not the growth of that body itself.

5). A subject of ongoing interest is the various triggering factors which have been proposed for large silicic systems. Saturation of the magma with volatiles is considered or
demonstrated to be the norm in large silicic systems (e.g. Dunbar & Hervig, 1992b; Stix & Layne, 1996; Wallace *et al.*, 1999; Liu *et al.*, 2006; Chesner & Luhr, 2010; Fowler & Spera, 2010), so that volatile saturation in itself cannot be a sole trigger for eruption onset (cf. Blake, 1984). Although intrusion of a more mafic magma, or buoyancy-driven overpressure of such a large volume of magma are often suggested as critical triggers for eruption (e.g. Sparks *et al.*, 1977; Caricchi *et al.*, 2014; Malfait *et al.*, 2014), recent studies have highlighted a role for external tectonic forces in both triggering and halting supereruptions (e.g. Allan *et al.*, 2012; Cooper *et al.*, 2012). Although our work has not found evidence in the crystal-specific record to support or disprove this hypothesis for the Bishop Tuff, it would appear that the longer timescales of mixing with the ‘bright-rim’ magma (Chamberlain *et al.*, 2014a, b) are too extended to represent a direct trigger as has been suggested by Wark *et al.* (2007). However, recent volcanism in Long Valley has been shown to be closely linked to faulting (e.g. Bursik *et al.*, 2003), and therefore it seems plausible that tectonic processes may have allowed or aided rise of the ‘bright-rim’ magma and contributed to triggering of the Bishop Tuff eruption itself. Whether the paleotectonic record could be probed in enough detail to investigate this hypothesis is unknown, but it represents a possible avenue for further exploration in the Long Valley region.

**CONCLUSIONS**

We reach four main conclusions relating to the pre-eruptive Bishop Tuff magma chamber and its relationship to rejuvenating magmatism.

1. Our data confirm that there was a single, unitary Bishop Tuff magma chamber that was stratified in temperature and composition even prior to intrusion of the ‘bright-rim’ magma. Available evidence strongly counts against proposals for the presence of two
laterally separated magma chambers at similar depths and temperatures feeding the early and late Bishop Tuff, respectively.

2. The stratification of the Bishop magma chamber was gradational from the upper parts (contributing to earlier-erupted deposits) yielding the most-evolved glass compositions, unzoned phenocrysts, pyroxene-free and lower model temperatures through to lower parts (dominating late-erupted deposits) with less-evolved glass compositions, systematically zoned sanidine and quartz phenocrysts, pyroxene-bearing and higher model temperatures (Figs. 4-8). There is an intervening transition zone between the dominant ~2/3 volume of more-evolved and the subordinate ~1/3 volume of less-evolved compositions. The transitional magma is represented in samples from the Sherwin lobe of Ig1Eb and Ig2Na which show mineralogical and geochemical characteristics in common with both earlier (eastern) and later (northern) erupted compositions.

3. We find no evidence that significant amounts of eastern-type magma were ever present prior to the eruption in the chamber below the northern vents of the Bishop eruption (Fig. 14). Similarly, there is a sparseness of x-ray less-evolved magma vented from the southerly and easterly parts of the caldera. We thus infer that there was an asymmetric roof and floor to the magma body, both stepped down to the north (Fig. 14), the former as previously proposed from melt inclusion studies.

4. From their significant enrichment in Ba, Sr, LREE and Ti (Fig. 9), the sparse dacitic clasts and dark pumices are confirmed as possible analogues for the ‘bright-rim’ magma that intruded the lower regions of the Bishop magma chamber. The ‘bright-rim’ magma cannot be represented by a single composition, however, and instead likely represents varied tappings of a chemically heterogeneous source. Its interaction with the pre-existing Bishop magma body was prolonged over hundreds to thousands of years, and affected
through mixing the lower ~1/3 by volume of an otherwise stably stratified and non-
convecting body of crystal-poor rhyolite

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**FIGURES AND CAPTIONS**

**Fig. 1.** Map of the Long Valley area, eastern California, USA, after Hildreth & Wilson (2007). The approximate initial vent location is after Hildreth & Mahood (1986).

Regions of present-day outcrops of Bishop ignimbrite are highlighted in dark grey. The envelope enclosing precursory Glass Mountain (GM) lava domes is shown in light grey. The line marked ‘Fall deposition envelope’ marks the westerly limits of Bishop fall deposits found in the proximal area shown in this map.
Fig. 2. Summary of the stratigraphy of the Bishop Tuff, after Hildreth & Wilson (2007).

Ignimbrite units are coloured: purple for the early-erupted units; green for the late-erupted units, with subordinate units in orange and pink. Fall units are in grey. Panel (a) shows a hypothetical cross section from NW to NE through the northern ignimbrite lobes as viewed from the centre of Long Valley caldera. Panel (b) is a schematic proximal to distal cross-section approximately along the line of Owens River Gorge, where the horizontal axis represents distance from the initial vent site of Hildreth & Mahood (1986). GM refers to Glass Mountain.
**Fig. 3.** Representative images to show the examples of textural variation observed in the common mineral phases found in Bishop pumices where sample numbers in italics relate to those in Table 1 (see Chamberlain et al., 2014a, for zircon zonation patterns). Panel (a) shows the different types of sanidine zoning identified from BSE imagery with Ba concentrations from cores and rims labelled in red, and panel (b) shows the normally unzoned plagioclase, and the subordinate oscillatory zoned plagioclase (BSE images). Panels (c) and (d) show the variations observed in BSE images of pyroxenes ranging from unzoned through to complexly zoned, where a brighter BSE shade in the rim represents higher Fe (and lower Mg) content, and a darker shade represents lower Fe (and higher Mg) content compared with the core. In all images the scale bar is 100 µm in length.
Fig. 4. Major (a) and trace (b to f) element variations in sanidines. Purple filled triangles represent sanidine analyses from ‘normal’ eastern samples; green filled circles ‘normal’ northern samples and pink filled squares are analyses from pumices in Ig1NW, Ig1SW and Ig2SW. Xenocrystic sanidines are outlined in orange on panel (a) but are not shown in any trace element plots. (a) Major element variation in Bishop sanidine; (b) Histogram of Ba (ppm) for all analyses of sanidine from normal eastern-
and northern-sourced samples; (c, e) Trace element variation in Bishop sanidine; (d) Histogram of Ba (ppm) for sanidine cores (darker) and rims (lighter) from eastern-(purple) and northern-sourced (green) ‘normal’ Bishop samples; (f) comparison of all normal Bishop sanidine analyses (eastern- and northern-sourced) with sanidine from variant (swirly and dacitic) pumice types. The approximate range in ‘bright rim’ Ba concentrations are shown on (d) by the arrow. The full data sets (and uncertainties) are given in Electronic Appendices 1 and 2.
Fig. 5. Major (a) and trace (b, c, d) element variations in plagioclase from Bishop Tuff samples, where (d) is an enlargement of (b) to show the overlap in composition of lower-Ca plagioclase from both northern and eastern samples. Purple filled triangles represent eastern material; green filled circles northern material. The higher-Ca plagioclase group is outlined by the blue field. The full data sets (and uncertainties) are given in Electronic Appendices 1 and 2.
Fig. 6. Biotite major element variations from Bishop Tuff samples. Panels (a) and (b) highlight the influence of alteration within the Bishop samples, especially those from the eastern Bishop Tuff. For comparison, the data of Hildreth (1979) are plotted as grey diamonds. The full data sets (and uncertainties) are given in Electronic Appendices 1 and 2.
Fig. 7. Major (a, b) and trace (c, d) element compositions of pyroxenes from northern units of the Bishop Tuff. Orthopyroxene (OPX) analyses are presented in the left column (panels a, c); clinopyroxene (CPX) in the right column (panels b, d). Cores are plotted as filled squares, rims as filled triangles. The full data sets (and uncertainties) are given in Electronic Appendices 1 and 2.
Fig. 8. Major (a) and trace (b to f) element variations in glass from samples in the crystal-poor to crystal-rich range of ‘normal’ pumices (as defined by Hildreth & Wilson, 2007) of the Bishop Tuff. Eastern samples are plotted as purple filled triangles (or purple fields); northern samples as green filled circles (or green fields). Ig1NW, Ig1SW and Ig2SW data are plotted as pink filled squares in panel (b); Ig2Na data as orange filled squares in panel (e). Illustrative representative single sample variations
are shown in panels (c) and (d) – but note the contrast in scales – with the range in all eastern (purple) and northern (green) samples demonstrated by the fields plotted in (c), (d) and (f). The full data sets (and uncertainties) are given in Electronic Appendices 1 and 2.
Fig. 9. Major (a) and trace (b to d) element variations in glass from samples in the range of ‘variant’ pumices (as defined by Hildreth & Wilson, 2007) in the Bishop Tuff compared with those measured from all ‘normal’ Bishop pumices (grey field): swirly pumices are red crosses, dacitic pumices are blue symbols, where lighter crosses are the dacitic glass in the dacite samples, and darker diamonds are the rhyolitic glass in the dacite samples. The dacite data plotted includes both the white admixed rhyolitic portions (as blue diamonds) and darker, dacitic portions (as blue crosses) of the dacite clasts. The full data sets (and uncertainties) are given in Electronic Appendices 1 and 2.
Fig. 10. Summary of crystal core compositional data providing evidence for thermal and compositional stratification within the Bishop Tuff prior to introduction of the ‘bright-rim’ magma. (a) Two-feldspar temperatures from Chamberlain et al. (2014b) plotted against the respective Ba content of sanidine cores from which the temperature was calculated. Eastern samples are purple filled triangles; northern samples are green filled circles. Panel (b) shows trace element variations in plagioclase cores: eastern samples are purple filled triangles; northern samples are green filled circles. Panel (c) shows the averages (as single points) and ranges (as bars) of trace element
concentrations in sanidine cores ordered by stratigraphic position and eruptive
ordering, where green triangles refer to Ba data, and red squares refer to Sr data.
Fig. 11. Modelling of inferred melt compositions from (a) high-Ca plagioclase (b) clinopyroxene and (c) orthopyroxene. The field of all Bishop glass data is shown in light grey, the range in melt inclusion compositions from Roberge et al. (2013) in dark grey, and the field for glass compositions of variant swirly pumices in purple. Dacite glass fields are shown in blue. Modelled glass compositions from clinopyroxene cores are shown as orange points, using the partition coefficients of Olin & Wolff (2010). Modelled glass compositions from higher-Ca plagioclase cores (yellow points) and from orthopyroxene cores (green points) are calculated from the multiple partition coefficients given by Brophy et al. (2011). Different shaped symbols in panels (a) and (b) indicate modelled melt compositions from the various partition coefficients calculated with each respective study.
Fig. 12. Cartoon of the Bishop magma chamber showing the inferred origins for the crystal species (including those native to the late-stage incoming ‘bright-rim’ magma or dacites) and their compositional variations found in different eruptive units of the Bishop Tuff. Examples of eruptive units (from Hildreth & Wilson, 2007) are detailed in bold, with specific representative sample numbers from this study (listed in Table 1) in italics. Variant pumice types are not shown for simplicity. See text for discussion.
Fig. 13. Plot of mean values and total ranges of variation in key trace element concentrations in glass of the Bishop Tuff ordered by stratigraphic unit and eruptive ordering.
**Fig. 14.** A revised model for the immediately pre-eruptive Bishop magma chamber derived from this work and utilising the depth estimates of Wallace et al. (1999). Black sills and dykes represent less-evolved compositions than the blue sills and dykes which represent the ‘bright rim’ magma. See text and Fig. 12 for discussion of various crystal origins.
Fig. 15. Comparison of crystal-poor (xp) and crystal-rich (xr) samples from package Ig2Na, showing (a) glass and (b) sanidine trace element concentrations compared with the ranges of concentrations from all eastern (purple shaded area) and northern (green shaded area) samples. Analyses from the xp sample (purple filled diamonds) of Ig2Na should overlap with the purple eastern field if it represented ‘eastern-type’ magma that was present in the part of the chamber tapped by the northern vents. Analyses from crystal rich samples are plotted as green triangles. The full data sets (and uncertainties) are given in Electronic Appendices 1 and 2.