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# Tectonic and climate history influence the geochemistry of large-volume silicic magmas: New $\delta^{18}O$ data from the Central Andes with comparison to N America and Kamchatka

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ABSTRACT

New $\delta^{18}O$ data from magmatic quartz, plagioclase and zircon crystals in Neogene large-volume, rhyodacitic ignimbrites from the Central Andean Ignimbrite Province reveal uniformly high-$\delta^{18}O$ values ($\delta^{18}O_{(Qtz)}$ from +8.1 to +9.6‰ − 43 analyses from 15 ignimbrites; $\delta^{18}O_{(Plag)}$ from +7.4 to +8.3‰ − 10 analyses from 6 ignimbrites; $\delta^{18}O_{(Zrc)}$ from +6.7 to +7.8‰ − 5 analyses from 4 ignimbrites). These data, combined with crustal radiogenic isotopic signatures of Sr, Nd and Pb, imply progressive contamination of basaltic magmas with up to 50 vol.% upper crust in these large volume silicic systems. The narrow range of $\delta^{18}O$ values also demonstrate that surprising homogeneity was achieved through space (100's km) and time (~10 Ma to recent) in these large-volume magmas, via residence in their parental middle to upper crustal bodies. Low-$\delta^{18}O$ values of many large volume (>10 km$^3$) silicic magmas in North America and Kamchatka, discussed here for comparison, reflect the influence of meteoric-hydrothermal events and glaciations in lowering these $\delta^{18}O$ values via the assimilation of hydrothermally-altered crustal material. Conversely, there is a scarcity of a low-$\delta^{18}O$ signature in the Central Andes and subduction-related or influenced systems in North America, such as the Oligocene Great Basin of Nevada and Utah, the Southern Rocky Mountain Volcanic Field of Colorado, and the SW Nevada volcanic field system. In these regions, the generally heavy-$\delta^{18}O$ magmatic signature is interpreted as a reflection of how a broadly compressional regime, high elevation, aridity and evaporation rates limit availability and infiltration of large amounts of surface meteoric water and hydrothermal alteration of the shallow crust. This leads us to speculate that the $\delta^{18}O$ values of large volume silicic magmas in these areas record a paleoelevation and paleoclimate signal. If this is the case, $\delta^{18}O$ values of ignimbrites can potentially be used to track the effects of a meteoric-hydrothermal derived $\delta^{18}O$ signature from upper crustal rocks that are subsequently assimilated to produce these magma types.

## 1. Introduction

It is often recognized that large volume silicic magma systems result from, and are driven by mantle-derived basalt (e.g., Hildreth, 1981). Volcanism associated with these systems is commonly referred to as 'supervolcanic' due to the prodigious volumes (>400 km$^3$) of individual eruptions. Thermal considerations for growing eruptible magma bodies of such size implicate transient excursions of elevated mantle flux and attendant thermal and mechanical preparation of the crust (de Silva and Gosnold, 2007; de Silva, 2008; Annen, 2009). Assimilation of significant volumes of crustal material is required to explain the compositions of the isotopically enriched erupted silicic magmas (e.g., Lipman et al., 1978; de Silva, 1989a; Francis et al., 1989; Bindeman, 2008). These systems therefore provide a valuable proxy for the age, composition, and structure of the crust being assimilated. However, other factors such as local processing of hydrothermally-altered crust also impact regional differences in isotopic composition, making mass balance estimations of different crust and mantle contributions challenging.

Since oxygen is the major element in rocks and magmas it can provide key constraints on source and crust mixing (e.g., Bindeman, 2008 and Refs therein). For example, silicic magmas erupted at active continental margins are often highly $\delta^{18}O$-enriched (up to +11‰) compared to mantle values (+5.7 ± 0.3‰), a difference that implicates significant melting and assimilation of high-$\delta^{18}O$ supracrustal sedimentary materials and igneous and metamorphic rocks subsequently derived from them (Taylor, 1980; James, 1981). $\delta^{18}O$ values are also sensitive to variations in the source regions of magma generation and the assimilation of low-$\delta^{18}O$ hydrothermally-altered

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shallow crustal rocks (e.g., Bindeman, 2008). Any variations in the contribution from these factors will produce magmas with differing $\delta^{18}$O values, especially for magmas from a long-lived caldera system. Such systems can also show temporal decreases in $\delta^{18}$O values related to progressive assimilation of successively greater amounts of low-$\delta^{18}$O hydrothermally-altered rock (e.g., Watts et al., 2011).

Here, we present new and published radiogenic isotope data from large volume silicic late Cenozoic ignimbrites from the well-known Neogene Central Andean Ignimbrite Province (de Silva et al., 2006), the surface manifestation of a subduction-related magmatic 'flare-up'. These data show that the Central Andean ignimbrite flare-up produced temporally and spatially homogeneous magmas with enriched $\delta^{18}$O values (>8‰), that share a general family resemblance to 'monotonous intermediate' magmas produced during the ignimbrite flare-ups in the Great Basin and the Southern Rocky Mountain Volcanic Field of the North American Cordillera. However, in comparison with many similar systems in North America and Kamchatka, we find no evidence for depleted $\delta^{18}$O values. Herein, we discuss the 'enriched' $\delta^{18}$O character of the large Central Andean ignimbrites in terms of their petrogenesis and then explore factors that may account for this enrichment. We deduce that the tectonic history (elevation) and climate (aridity) have influenced the $\delta^{18}$O character of Neogene Central Andean supervolcanic systems and find a corollary in the Oligocene Great Basin ('Nevadaplano') of the North American Cordillera.

## 2. Geological and geochemical background to Central Andean large silicic magmatic systems

Some of the best preserved large scale silicic volcanic fields on Earth are located in the Central Andes, where hyper-aridity and high elevation combine to preserve extensive Neogene ignimbrites and their source calderas (de Silva and Francis, 1991). Two of the best studied regions are the 70,000 $km^2$, 10 Ma to recent Altiplano Puna Volcanic Complex (APVC; 21° and 24° S) where >13,000 $km^3$ of silicic magmas were erupted episodically (de Silva and Gosnold, 2007; Salisbury et al., 2011), and the 6 to 2 Ma Cerro Galán system at ~26° S, where >1200 $km^3$ of magma was erupted (Sparks et al., 1985; Folkes et al., 2011c). These are thought to be part of a regional magmatic ignimbrite flare-up that is related to the tectonic evolution of the Central Andes (de Silva, 1989b; Kay and Kay, 1993; Allmendinger et al., 1997; de Silva et al., 2006; Gleeson et al., 2011). Key factors driving the flare-up include crustal shortening and thickening between 25 and 10 Ma that produced crustal thicknesses of 60–80 km throughout the Central Andes (e.g., Isacks, 1988), and subsequent delamination of the weakened subcontinental mantle lithosphere (see Kay et al., 1999) that facilitated the transient high mantle power input required to drive large volume silicic magmatism (Hildreth and Moorbath, 1988; de Silva and Gosnold, 2007). Located on a highland plateau with an average elevation of ~4000 m and individual summits over 6000 m (Fig. 1A), the APVC and Cerro Galán are examples of integrated (multicaldera) and individual silicic magmatic systems (respectively). These, like similar systems around the world, represent the surface expression of episodically-constructed batholiths over time scales of ~10 Ma and ~4 Ma respectively (de Silva and Gosnold, 2007; Folkes et al., 2011c; Salisbury et al., 2011). However, despite such high elevation, these regions are largely ice-free.

The large ignimbrites of the APVC and Cerro Galán are large 'monotonous intermediates' following the nomenclature of Hildreth (1981); they are crystal-rich (>40% crystals), calc-alkaline, high-K dacites to rhyodacites. The mineralogy of these ignimbrites consists of varying proportions of plagioclase, biotite, quartz, amphibole, Fe–Ti oxides and minor sanidine, apatite, zircon, titanite, monazite, and allanite. Volatile ($H_2O$ and $CO_2$) solubility calculations from melt inclusion data and Al-in-hornblende barometry indicate shallow chambers (tops between 4 and 6 km depth) and magmatic temperatures from 780 to 830 °C prior to eruption (Schmitt, 2001; Kay and Coira, 2009). The ignimbrites all exhibit typical arc affinities with high Ba/Nb and Ba/La ratios and enrichment in the light rare earth elements. Detailed geochemical and petrological studies have shown a general chemical homogeneity in bulk geochemical properties both through time (Cerro Galán, Folkes et al., 2011b) and through time *and* space (the APVC, Schmitt et al., 2001; de Silva and Gosnold, 2007).

## 3. Methods

Oxygen isotope analyses were performed at the University of Oregon stable isotope laboratory using $CO_2$-laser fluorination and a 35 W laser. Individual and bulk mineral grains ranging in weight between 0.6 and 2 mg were reacted in the presence of purified $BrF_5$ reagent to liberate oxygen. The gases generated in the laser chamber were purified through a series of cryogenic traps held at liquid nitrogen temperature, and a mercury diffusion pump to eliminate traces of

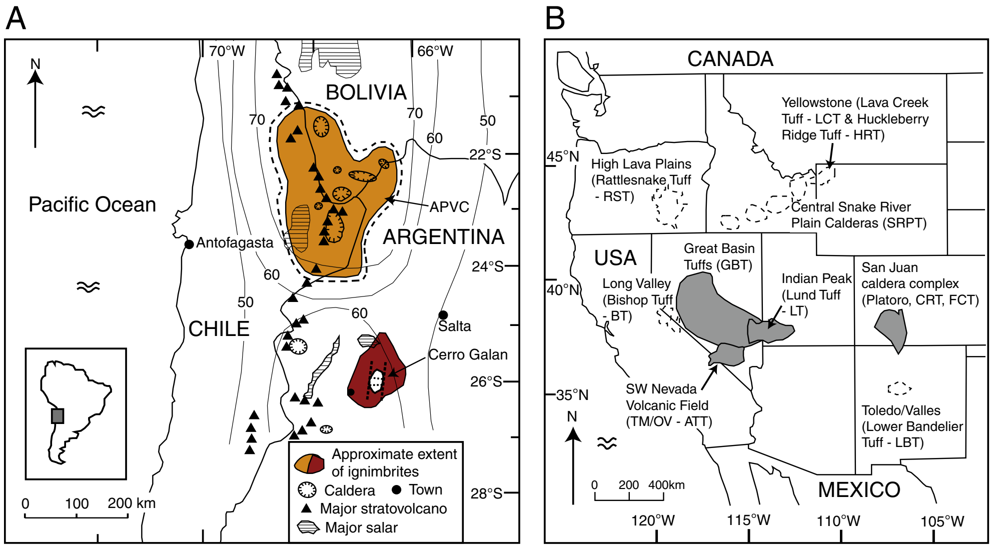

**Fig. 1.** (A) Locations of the major silicic calderas described in the Central Andes (modified from Folkes et al., 2011c). Contour lines show crustal thicknesses in km since 10 Ma (Wigger et al., 1994). (B) Map of the western US showing the locations and approximate distributions of the large-volume tuffs (and calderas of the Snake River Plain–Yellowstone systems) described in this study. Solid (filled) fields are subduction related or influenced systems; other systems (dashed areas) are shown for comparison purposes only.

fluorine gas. Oxygen was converted to $CO_2$ gas by a small platinum-graphite converter, the yield was measured, and then $CO_2$ gas was analyzed on a MAT 253 mass spectrometer. Three to seven standards were analyzed together with the unknowns during each analytical session. San Carlos olivine ($\delta^{18}O = +5.35‰$) and Gore Mt garnet ($\delta^{18}O = +5.75‰$) were used as standards. Day-to-day $\delta^{18}O$ variability ($= \delta^{18}O_{(Standard)} - \delta^{18}O_{(Measured\ standard)}$) ranged from 0.1 to 0.25‰ and these values were added to the unknowns to correct for day to day standard offset. Values are reported on Standard Mean Ocean Water (SMOW) scale. The precision on standards and duplicates of individual quartz, feldspar and zircon analyses is typically better than 0.1‰.

**Table 1**
New $\delta^{18}O$ data for Cerro Galan and APVC ignimbrites.

| Ignimbrite sample[a]<br>(eruption age, Ma)[b]<br>Cerro Galan | $\delta^{18}O$<br>(‰) | Ignimbrite sample[a]<br>(eruption age, Ma)[c]<br>APVC | $\delta^{18}O$<br>(‰) |
|---|---|---|---|
| Blanco (N.D.)[d] | | Artola (9.40 ± 0.03) | |
| *CG 475* | *9.25 ± 0.065* | 83004B | 8.10 ± 0.17 |
| *CG 475* | *9.12 ± 0.065* | 83004B | 8.64 ± 0.17 |
| Lower Merihuaca (5.60 ± 0.20) | | Guacha (5.65 ± 0.01) | |
| *CG 476* | *9.59 ± 0.06* | 88022 | 8.31 ± 0.03 |
| *CG 476* | *9.47 ± 0.06* | 88022 | 8.29 ± 0.03 |
| | | 88022 | 8.37 ± 0.03 |
| Middle Merihuaca (5.56 ± 0.10) | | Puripicar (4.09 ± 0.02) | |
| CG 80 | 8.82 ± 0.025 | 83018 | 9.62 ± 0.04 |
| CG 80 | 8.93 ± 0.025 | 83018 | 9.54 ± 0.04 |
| *CG 80* | *8.77 ± 0.025* | 83021 | 9.44 ± 0.18 |
| CG 83 (Plag) | 7.74 ± 0.025 | 83021 | 9.09 ± 0.18 |
| CG 83 (Plag) | 7.79 ± 0.025 | | |
| | | Toconao (4.5–4.0) | |
| Upper Merihuaca (5.49 ± 0.11) | | 88011 (Plag) | 7.84 ± 0.14 |
| *CG 91* | *9.14 ± 0.02* | 88011 (Plag) | 7.57 ± 0.14 |
| CG 91 (Plag) | 8.25 ± 0.02 | | |
| CG 91 | 9.18 ± 0.02 | Atana (3.96 ± 0.02) | |
| | | 88053 | 8.37 ± 0.04 |
| | | 88053 | 8.29 ± 0.04 |
| Pitas (4.84 ± 0.04) | | 88055 | 8.54 ± 0.08 |
| *CG 465* | *9.05 ± 0.11* | 88055 (Plag) | 7.37 ± 0.08 |
| *CG 465* | *8.83 ± 0.11* | 88055 (Plag) | 7.54 ± 0.08 |
| *CG 461 (Zircon)* | *6.65* | 88056 | 8.28 ± 0.01 |
| | | 88056 | 8.26 ± 0.01 |
| Real Grande (4.68 ± 0.07) | | | |
| *CG 112* | *9.04 ± 0.11* | | |
| *CG 112* | *8.82 ± 0.11* | | |
| CG 112 | 9.44 ± 0.11 | Tara (3.49 ± 0.01) | |
| *CG 106 (Zircon)* | *7.67* | 88047 | 8.31 ± 0.14 |
| | | 88047 | 8.59 ± 0.14 |
| Cueva Negra (3.78 ± 0.08) | | 88047 (Plag) | 7.66 ± 0.14 |
| *CG 436* | *9.37 ± 0.10* | 88023 | 8.41 ± 0.11 |
| *CG 436* | *9.57 ± 0.10* | 88023 | 8.37 ± 0.11 |
| | | 88023 | 8.71 ± 0.11 |
| Pre-CGI ignimbrite (2.80 ± 0.04) | | | |
| *CG 530* | *9.47 ± 0.02* | | |
| *CG 530* | *9.51 ± 0.02* | Purico (1.3–1.0) | |
| | | 83041 | 8.47 ± 0.08 |
| CGI (2.08 ± 0.11) | | 83041 (Plag) | 7.53 ± 0.03 |
| CG 141 | 9.04 ± 0.276 | 83041 (Plag) | 7.48 ± 0.03 |
| CG 141 | 9.04 ± 0.276 | *83041 (Zircon)* | *7.78* |
| CG 141 | 9.59 ± 0.13 | | |
| *CG 428* | *9.15 ± 0.13* | Average of all APVC quartz data | 8.60 ± 0.45 |
| *CG 428* | *9.41 ± 0.13* | | |
| *CG 27b (Zircon)* | *6.89* | | |
| *CG 456 (Zircon)* | *7.06* | | |
| Average of all Galan quartz data | 9.18 ± 0.28 | | |

[a] All analyses are of single quartz crystals except for the those in italics (bulk crystals) and single crystals labeled plag (plagioclase) or zircon.
[b] Eruption ages for the Galan Ignimbrites from $^{40}Ar/^{39}Ar$ analyses (Folkes et al., 2011c).
[c] APVC eruption ages from de Silva and Gosnold (2007) and Salisbury et al. (2011).
[d] N.D. = not determined.

## 4. Oxygen isotope data for the Central Andes

In total, 58 separate analyses (2 to 5 from each ignimbrite) were performed on single and bulk crystal separates of quartz from pumice from the Cerro Galán and APVC ignimbrites (Table 1). For some samples, plagioclase ($An_{40-50}$) and zircon crystals were also analyzed, yielding $\delta^{18}O_{(Qtz-Plag)}$ fractionation $\approx 1‰$, and $\delta^{18}O_{(Qtz-Zrc)}$ fractionation $\approx 2–2.5‰$, consistent with typical magmatic temperatures of 750 to 850 °C for these ignimbrites. For these magmas, the $\delta^{18}O_{(Qtz-melt)}$ fractionation is approximately 0.5‰ (Matthews et al., 1994; Bindeman and Valley, 2002).

$\delta^{18}O_{(Qtz)}$ values from the Galán ignimbrites are moderately high-$\delta^{18}O$ and vary over a relatively narrow range, from +8.77 to +9.59‰, in repeated eruptions over >3.5 Ma (Fig. 2A). The variation within each of the older Galán ignimbrites (~6 to 2.8 Ma) is less than 0.22‰ with overlapping errors (except the Real Grande Ignimbrite), while the ~2.1 Ma Cerro Galán Ignimbrite has a range of +9.04 to +9.59‰ that overlaps the values for all the other ignimbrites (Table 1; Fig. 2A). There is no correlation between $\delta^{18}O_{(Qtz)}$ values and age (Fig. 2A) in the whole Galán suite, nor is there a correlation between $\delta^{18}O_{(Qtz)}$ values and erupted volumes (Fig. 2B). The average $\delta^{18}O_{(Qtz)}$ value of the Galán data is +9.18 ± 0.28‰ (one standard deviation); the average $\delta^{18}O_{(Plag)}$ value is +7.93 ± 0.023‰; and the average $\delta^{18}O_{(Zrc)}$ value is +7.07‰, in magmatic equilibrium.

Some single crystals in APVC ignimbrites yield elevated $\delta^{18}O$ values. Quartz crystals in the Puripicar Ignimbrite (4.09 Ma, >500 km$^3$) yield the highest values at +9.09 to +9.62‰, while the Artola Ignimbrite (9.4 Ma, ~100 km$^3$) yields the lowest values of +8.10 to +8.64‰ (Table 1). In general, $\delta^{18}O_{(Qtz)}$ values in the APVC ignimbrites yield a lower average and wider range (+8.60 ± 0.45‰) than the Galán ignimbrites, with $\delta^{18}O_{(Qtz)}$ values for the latter plotting at the upper end of the APVC range (Table 1; Fig. 2).

Single crystal plagioclase and zircon crystals from the same ignimbrites show similar elevated $\delta^{18}O$ values to that of $\delta^{18}O_{(Qtz)}$. $\delta^{18}O_{(Plag)}$ values are restricted to +7.4 to +8.3‰, and $\delta^{18}O_{(Zrc)}$ values from +6.7 to +7.8‰ (Table 1). Where multiple samples have been analyzed for a single ignimbrite (e.g., the CGI and Middle Merihuaca ignimbrites at Cerro Galán and the Toconao, Atana and Purico ignimbrites in the APVC; Table 1), they yield values that overlap within error. It is also important to note that $\delta^{18}O_{(Plag)}$ values for the high-Si rhyolitic Toconao Ignimbrite (+7.57 to +7.84‰) are similar to the co-genetic Atana Ignimbrite (+7.37 to +7.54‰; both erupted from the La Pacana caldera ~4.0 Ma) and with all other $\delta^{18}O_{(Plag)}$ values from Cerro Galán and the APVC (Table 1).

## 5. Discussion

### 5.1. Chemical homogeneity in large silicic systems in the Central Andes

The new data we present shows remarkable $\delta^{18}O$ homogeneity at a variety of temporal and spatial scales:

1) The maximum range of $\delta^{18}O_{(Qtz)}$ values within a single ignimbrite in the Central Andes is seen in the Cerro Galán Ignimbrite (+9.04 to 9.59‰; Table 1). Additionally, our data shows that early and late erupted ignimbrites (from the same cooling unit) are homogeneous in $\delta^{18}O$ values. This indicates that the Central Andean supervolcanic systems exhibit isotope homogeneity within a single eruptive unit, confirming them as ‘isotopic monotonous intermediates’ in the sense of Bindeman and Valley (2002).
2) Successive eruptions from long-lived confocal magmatic systems are isotopically homogeneous. $\delta^{18}O_{(Qtz)}$ values from the Galán system ranges from 8.83 ± 0.11 to 9.59 ± 0.06‰ (Table 1) over a 6 Ma period of confocal volcanism. Similarly, the 5.65 to 3.49 Ma Guacha–Tara Caldera Complex, and the adjacent ~4 to 1 Ma La Pacana–Purico Volcanic Complex show a virtually unresolvable

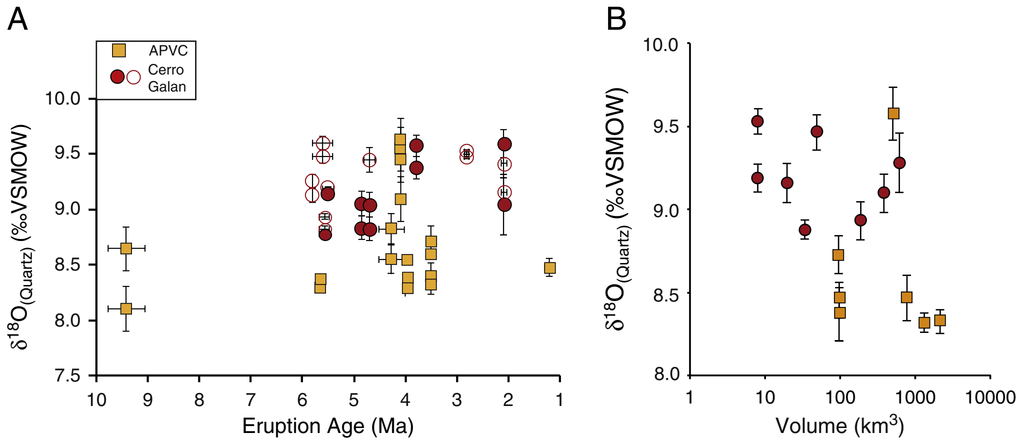

**Fig. 2.** (A) $\delta^{18}O_{(Qtz)}$ values of Cerro Galán and APVC ignimbrites. Note that the $\delta^{18}O_{(Qtz)}$ values for the Toconao ignimbrite have been estimated by adding 1.0‰ to the $\delta^{18}O_{(Plag)}$ analyses (see text for rationale). Open symbols = bulk crystals, closed symbols = single crystals. (B) $\delta^{18}O_{(Qtz)}$ vs. ignimbrite volume for central Andean samples. The average $\delta^{18}O_{(Qtz)}$ value for each ignimbrite is shown.

range in $\delta^{18}O$ values with time (Table 1). This attests to the local isotopic homogeneity throughout the magmatic lifetime of these systems.

3) The ~4 Ma crystal-poor, high-Si rhyolite and the crystal-rich dacite pair of the Toconao and Atana ignimbrites (respectively) from the La Pacana caldera (Lindsay et al., 2001; Schmitt et al., 2002) show a range in $\delta^{18}O_{(melt)}$ values of 8.11 to 7.83 ± 0.04‰, respectively. The data suggest that even individual 'zoned' magmatic systems in the central Andes are isotopically homogeneous.
4) The similarity of $\delta^{18}O$ values during at least 10 Ma of silicic magmatism over a spatial scale of several hundreds of kilometers, suggest regional spatial and temporal homogeneity in magmatic oxygen isotopes during the Neogene ignimbrite flare-up.

The most obvious systemic signal of our single crystal oxygen-isotope data from the Neogene ignimbrite flare-up in the Central Andes is the elevated but restricted $\delta^{18}O$ range. This is in contrast to previous studies (Harmon et al., 1984; Francis et al., 1989) that used whole rock bulk matrix material and pumice from these same ignimbrites to report a much broader and scattered range of $\delta^{18}O$ values from +7 to +14‰. This newly-demonstrated trans-temporal, -spatial, -compositional, and -volumetric isotopic homogeneity implies consistent magma generation, fractionation and crustal assimilation histories operating over ~10 Ma. This is quite remarkable given the likely $\delta^{18}O$ (and $^{87}Sr/^{86}Sr$, $^{143}Nd/^{144}Nd$ and $^{206}Pb/^{204}Pb$) heterogeneity of the putative crustal source rocks in the Central Andes (Taylor, 1980; Lucassen et al., 2001). The Cerro Galán and APVC ignimbrites overlap in $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ (Figs. 3A,B) space, but variations in these values between the two regions probably reflect varying proportions of Precambrian crust. We treat them as representative of the Neogene ignimbrite flare-up. Furthermore, this $\delta^{18}O$ homogeneity is present throughout the entire $SiO_2$ range of large ignimbrites sampled in the Central Andes; the crystal-rich dacites and rhyodacites have similar, elevated $\delta^{18}O$ values to the Toconao ignimbrite; a crystal-poor high-$SiO_2$ rhyolite. This suggests that variations in magmatic whole-rock geochemistry and pre-eruptive processes did not affect the underlying high-$\delta^{18}O$ values of large volume magmas generated in the Central Andes.

### *5.2. Sr–O–Nd correlations and implications for magma-genesis in the Central Andes*

Petrogenetic and thermal models for large scale silicic magmatism and volcanism in the Central Andes constrained by geophysical observations connote elevated power input from the mantle that results in progressive assimilation-fractional crystallization processes at different levels of the crust (de Silva et al., 2006 and references therein). Thus, while true primitive magmas were apparently not erupted

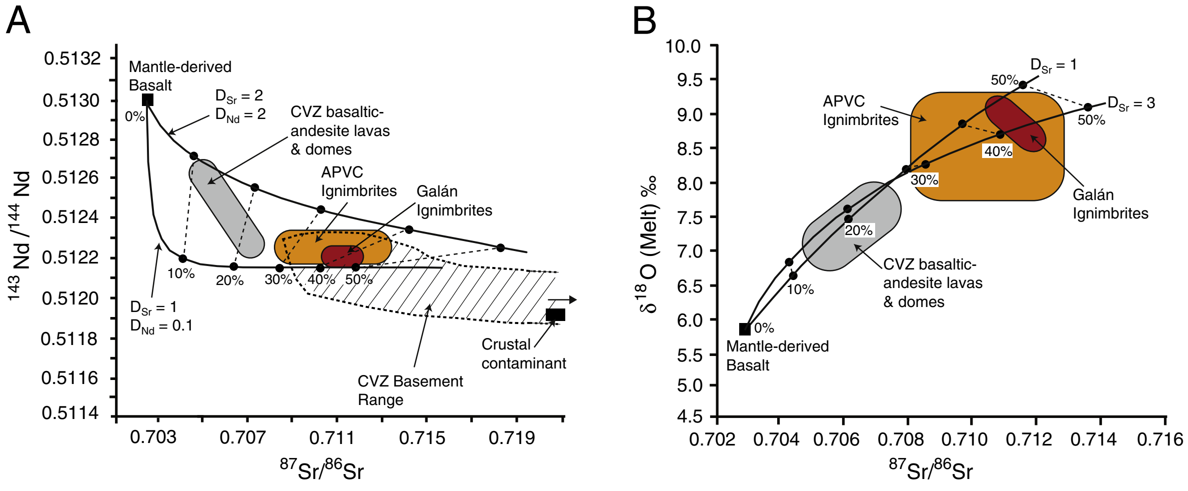

**Fig. 3.** Sr–Nd–O isotopic data for large volume silicic magmas and smaller-volume basaltic andesite lavas in the Central Andes. Simple mixing models are shown, using EC-$RA_XFC$ modeling (Bohrson and Spera, 2007) with progressive amounts of crustal mixing (shown in %) with a basaltic magma. Initial mantle-derived basalt composition from Schmitt et al. (2001), crustal assimilant composition from Ort et al. (1996) and CVZ basement range from Lucassen et al. (2001).

in the Neogene Central Andes (e.g., Davidson, 1991), it is generally recognized that the silicic magmas are fundamentally derived from mantle-derived basalts that undergo extensive multi-level crustal filtering (e.g., de Silva et al., 2006), with lower crustal MASH (Mixing, Assimilation, Storage, and Homogenization; Hildreth and Moorbath, 1988) processes imparting enriched baseline compositions of $^{87}Sr/^{86}Sr > 0.705$ to Central Andean magmas (Davidson et al., 1991). In this context, our newly measured $\delta^{18}O$ values and available whole rock Sr and Nd isotopic data can be used to constrain the mass balance of crust and mantle during magmagenesis in the Central Andes (Fig. 3).

Petrogenetic modeling of magmatic isotope values is sensitive to the compositions selected for the parent magma and crustal contaminants. Kay et al. (2010) provide a useful overview of the different values used for these reservoirs in various petrogenetic modeling studies of magmas in the Central Andes, and the authors direct the reader there for a detailed discussion. The style of crustal contamination also needs to be considered. Recently, McLeod et al. (2012) showed that contamination of Central Andean parental magmas may proceed through interaction with small-volume upper crustal anatectic melts. While this is likely an important process for smaller 'steady-state' systems (composite cones and small volumetric centers) of the modern arc, bulk assimilation of crust is more likely during large volume 'flare-up' magmatism.

In this study, we chose to use the isotopic values of Schmitt et al. (2001) in their study of the Purico and La Pacana systems for a starting composition ($^{87}Sr/^{86}Sr = 0.7028$, $^{143}Nd/^{144}Nd = 0.5130$, and $\delta^{18}O_{(Magma)} = +5.8$‰; Fig. 3A,B) as the initial mantle-derived basaltic melts delivered to the lower crust of the Central Andes. The bulk composition of the crustal contaminant ($^{87}Sr/^{86}Sr = 0.7350$, $^{143}Nd/^{144}Nd = 0.5119$ and $\delta^{18}O_{(Magma)} = +12.5$‰) used in our modeling was taken from Ort et al. (1996) from a basement xenolith found in the Panizos ignimbrite. These values are within the range of likely basement material (e.g., orthogneiss, paragneiss, and granitoid xenoliths and lithic clasts) analyzed by Lucassen et al. (2001) and within the range of the crustal contaminants selected for petrogenetic modeling by Kay et al. (2010). A range of bulk partition coefficients for Sr ($D_{Sr} = 1$ to 3) and Nd ($D_{Nd} = 0.1$ to 2) were used in modeling calculations that reflect the likely maximum range of partitioning of these trace elements in the magmatic systems of the Central Andes.

We used the Energy Constrained Recharge, Assimilation, and Fractionation (EC-RA$_X$FC) mixing models of Bohrson and Spera (2007) to model whether the isotopic signatures of the primitive basaltic andesite lavas and the large volume silicic ignimbrites can be produced by contamination of primitive basaltic magma with incremental additions of partial melts of the crustal compositions described above. Table 2 and Fig. 3 show a summary of the results of the modeling with different partition coefficients for Sr and Nd and the progressive assimilation of country rock partial melts. The modeling shows that the most primitive basaltic andesite lavas in the Central Andes require 15–25 vol.% crustal assimilation, while the large-volume silicic magmas require at least 30 to 50 vol.% crustal assimilation from the starting basaltic composition (Fig. 3). This is a similar amount to that calculated by Kay et al. (2011) for the Galán ignimbrites and Kay et al. (2010) for the APVC ignimbrites. Using the Nd isotopic values listed above, the calculated Neodymium Crustal Index (NCI; DePaolo et al., 1992) for the Galán ignimbrites is 0.61 to 0.72, overlapping with values for the APVC ignimbrites (0.49 to 0.83; Schmitt, 1999; Schmitt et al., 2001).

Our analysis broadly concurs with previous work going back 30 years and reflects the convergence of thought on the petrogenesis of these supervolcanic systems. While all models agree that large scale crustal assimilation is important, the location and timing of assimilation in the pre-eruptive history of magmas remain contentious (e.g., Davidson et al., 1991; Stern, 1991). Some studies have suggested that a mantle source enriched by the introduction of sediment via subduction or underthrusting may result in elevated $\delta^{18}O$ values in the source region (Kay et al., 2011). This notwithstanding, the $\delta^{18}O$ values of the large volume ignimbrites still require significant additions of high-$\delta^{18}O$ materials beyond any sub-arc mantle enrichment, and this signature has to largely come from the crust. The elevated $\delta^{18}O$ values (+6.9 to +8.5‰; James, 1982; Harmon et al., 1984) of the most primitive basaltic andesite lavas in the Central Andean region suggest that substantial amounts of high-$\delta^{18}O$ material was involved in lower crustal petrogenesis of these magmas — the aforementioned MASH process of Hildreth and Moorbath (1988). This inference is supported by recent studies of large volume magmas generated in the APVC and at Cerro Galán, suggesting that the underlying chemical signatures (and also chemical homogeneity) are imparted on these magmas before they are delivered to the middle and the upper crust (e.g., Folkes et al., 2011b; Kern et al., 2011). A lower to middle crustal MASH zone, similar to that of Hildreth and Moorbath (1988), is proposed to have given the Cerro Galán and APVC large volume magmas their underlying isotopic, and trace element signatures. Kay et al. (2010) show that all of the Central Andean large-volume ignimbrites exhibit steep heavy REE patterns (Sm/Yb = 2.5 to 8), indicating a lower to middle crust magmagenesis.

**Table 2**
EC-RAFC modeling results for APVC and Cerro Galan ignimbrites.

| | X Ma[a] | $\delta^{18}O$ | $^{87}Sr/^{86}Sr$ | $^{143}Nd/^{144}Nd$ |
|---|---|---|---|---|
| $D_{Sr} = 2$ | 0.00 | 5.80 | 0.7028 | 0.51300 |
| $D_{Nd} = 2$ | 0.10 | 6.65 | 0.7046 | 0.51277 |
| | 0.20 | 7.42 | 0.7073 | 0.51260 |
| | 0.30 | 8.05 | 0.7105 | 0.51248 |
| | 0.40 | 8.62 | 0.7144 | 0.51237 |
| | 0.50 | 9.07 | 0.7186 | 0.51228 |
| $D_{Sr} = 2$ | 0.00 | 5.80 | 0.7028 | 0.51300 |
| $D_{Nd} = 0.1$ | 0.10 | 6.65 | 0.7046 | 0.51221 |
| | 0.20 | 7.42 | 0.7073 | 0.51218 |
| | 0.30 | 8.05 | 0.7105 | 0.51217 |
| | 0.40 | 8.62 | 0.7144 | 0.51217 |
| | 0.50 | 9.07 | 0.7186 | 0.51217 |
| $D_{Sr} = 1$ | 0.00 | 5.80 | 0.7028 | 0.51300 |
| $D_{Nd} = 0.1$ | 0.10 | 6.65 | 0.7046 | 0.51221 |
| | 0.20 | 7.42 | 0.7064 | 0.51218 |
| | 0.30 | 8.05 | 0.7081 | 0.51217 |
| | 0.40 | 8.62 | 0.7099 | 0.51217 |
| | 0.50 | 9.07 | 0.7115 | 0.51217 |
| $D_{Sr} = 3$ | 0.00 | 5.80 | 0.7028 | 0.51300 |
| $D_{Nd} = 0.1$ | 0.10 | 6.65 | 0.7040 | 0.51221 |
| | 0.20 | 7.42 | 0.7060 | 0.51218 |
| | 0.30 | 8.05 | 0.7086 | 0.51217 |
| | 0.40 | 8.62 | 0.7123 | 0.51217 |
| | 0.50 | 9.07 | 0.7167 | 0.51217 |
| Starting magma[b]: | | 5.80 | 0.7028 | 0.51300 |
| Crustal assimilant[c]: | | 12.50 | 0.7350 | 0.51190 |

[a] X Ma is the normalized proportion of wall-rock (crust) partial melts incrementally assimilated (see Bohrson and Spera, 2007).

[b] Starting basaltic magma composition for all modeling calculations taken from Schmitt et al. (2001).

[c] Crustal assimilant composition for all modeling calculations taken from Ort et al. (1996).

However, the correlation between radiogenic isotopes of large ignimbrites and local basement in the APVC (e.g., de Silva, 1989a; Lindsay et al., 2001; Schmitt et al., 2001) and Cerro Galán (Francis et al., 1989) as well as elsewhere in the Central Andes (e.g., the Frailes plateau, Schneider, 1987) suggests that upper crustal assimilation also influences the isotopic character of the large volume silicic magmas. The presence of xenocrysts of zircon from local basement rocks and the calculated long residence times (several hundred thousand years, from U–Pb dating of zircon autocrysts) of large volume magmas in shallow storage zones, in both the APVC (Schmitt et al., 2002; Kern et al., 2011) and the Galán ignimbrites (Folkes et al., 2011a), supports the inference that upper crustal assimilation is an important contribution to the ignimbrites' isotopic characteristics.

### 5.3. Comparison with large silicic systems in the North America Cordillera

The North American Cordillera of the western US includes the major and diverse tectonomagmatic provinces of the Great Basin, Colorado Plateau, Sierra Nevada, Snake River Plain and Rio Grande Rift. These provinces individually and collectively are amongst the most intensely studied areas of large scale silicic volcanism on Earth (Fig. 1B). The Great Basin and Colorado Plateau (the Southern Rocky Mountain Volcanic Field — SRMVF) are representative of massive outpourings of silicic magma during the Mid-Tertiary ignimbrite flare-up (Coney, 1978) of similar magnitude to the Neogene flare-up in the Central Andes as a whole. In fact, the 38 to 16 Ma episode in the Great Basin of Nevada and Utah generated at least a dozen supervolcanic eruptions (Best et al., 2009 and references therein) including the 3000 km$^3$ Lund Tuff (Table 3), with associated intrusive magmas (Solomon and Taylor, 1989). This region also contains the SW Nevada Volcanic Field of ~16 to 9 Ma containing the Timber Mountain and Oasis Valley (TM/OV) caldera systems related to subduction of the Farallon Plate and subsequent regional extension (Bindeman et al., 2006 and references therein). Further to the east, the eastern margin of the broad North American Cordillera is the 38–23 Ma Southern Rocky Mountain Volcanic Field (SRMVF) containing the San Juan caldera system of SW Colorado (Lipman, 2007) with very similar spatiotemporal-volume relations to the APVC. Other multicaldera ‘regional’ systems of note are the Snake River Plain of Idaho (e.g., Bonnichsen, 1982; Pierce and Morgan, 1992) and its western ‘mirror image’ in the High Lava Plains of Oregon. Other well studied large-volume ignimbrite-producing magmatic systems in the western US, include the ~1 Ma Long Valley caldera system of California (e.g., Bailey et al., 1976), the 2 Ma Jemez volcanic field and the Valles/Toledo caldera system of New Mexico (Smith and Bailey, 1968; Self et al., 1986), and the <2 Ma Yellowstone caldera system in Wyoming (e.g., Christiansen, 2001), the latter being the youngest eruption phase in the Snake River Plain region.

General chemical and physical differences exist between ignimbrites from North America and the Central Andes. The majority of large volume ignimbrites in the Central Andes are crystal-rich (>30 vol.%) dacites to rhyodacites (64 to 71 wt.% $SiO_2$); crystal poor high-Si rhyolite is rare, the Toconao ignimbrite (77 wt.% $SiO_2$; Fig. 4D, E) being the only one of note. While several regions of North American tuffs are also dominated by crystal-rich dacites to rhyodacites (e.g., several ‘Nevadaplano’ and Colorado Plateau systems such as the Lund and Fish Canyon Tuffs as well as tuffs associated with the Platoro and other Oligocene Great Basin systems) there are many more examples of the high-Si, crystal-poor variety, such as those in Snake River Plain/Yellowstone, Long Valley and Valles (Fig. 4D,E), as well as units with peralkaline affinities (e.g., the TM/OV calderas).

Furthermore, unlike the Central Andean systems we have studied, the age and geologic setting of these North American systems are quite varied. While the Oligocene Great Basin, SRMVF, and the SW Nevada volcanic field systems are broadly compressional, subduction-related or a result of slab delamination (similar to the Neogene tectonic setting in the Central Andes), other North American systems are extensional or plume-related. The Central Snake River Plain and Yellowstone ignimbrites are thought to be the product of ~16 Ma of mantle plume interaction with the west-southwest moving continental North American plate. The Jemez Mountain Volcanic Field and the Valles/Toledo caldera system are part of the Rio Grande Rift, while the Long Valley caldera system that produced the Bishop Tuff is associated with a transtensional tectonic setting (Bursik and Sieh, 1989). Crustal thicknesses prior to, and synchronous with, the eruption of the large volume silicic magmas in North America are also highly variable (Fig. 4F). The Snake River Plain, Bishop and Rattlesnake Tuffs were erupted above crust that was relatively thin (30 to 45 km) during their respective eruptions, in contrast to the Oligocene Great Basin Tuffs in the east of the basin that were erupted onto much thicker crust (up to 70 km; Fig. 4F, Coney and Harms, 1984; Streck and Grunder, 1999; Best et al., 2009). In comparison, syn-eruptive crustal thicknesses for ignimbrites erupted in the Central Andes show a higher and more restricted range (55 to 70 km; Fig. 4F).

Although the North American cordilleran volcanic fields developed in a variety of tectonic settings, the thermomechanical environment for developing these magmatic systems was broadly similar to the Neogene flare-up in the Central Andes (e.g., Hildreth, 1981; Best and Christiansen, 1991). Moreover, like the Central Andes, the isotopic composition of erupted volumes of 100's of km$^3$ require significant proportions of radiogenic crustal material in their petrogenesis, and large-scale average/mixing of diverse crustal isotopic compositions (Johnson, 1991; Johnson, 1993). The systems are thus broadly comparable.

The variation in the geologic and tectonic settings of the North American ignimbrites and the consistency of those for the Central Andes are also reflected in their radiogenic (Sr, Nd and Pb) isotopic data. The Central Andean ignimbrites have restricted isotopic ratios ($^{87}Sr/^{86}Sr = 0.7080$ to 0.7130; $^{143}Nd/^{144}Nd = 0.5123$ to 0.5121; and $^{206}Pb/^{204}Pb = 18.8–19.0$; Fig. 4A–C). In contrast, the North American tuffs exhibit a wider range in $^{87}Sr/^{86}Sr$ (0.7040 to 0.7160; Fig. 4A), $^{143}Nd/^{144}Nd$ (0.5129 to 0.5116; Fig. 4A) and $^{206}Pb/^{204}Pb$ (17.0 to 19.3; Fig. 4C) ratios. Each region has a broadly consistent set of characteristics that covers the compositional spectrum of ignimbrites in a region, but regions are quite distinct from each other. These variations in isotopic ratios for the North American ignimbrites can be correlated with the disparate crustal types, ages and their compositions found across the western portion of North America. The crust in these regions varies in age from Archean gneisses and greenstones (>2.7 Ga) to Late Mesozoic granites and diorites (~80 Ma), with a wide spectrum of Nd ($\varepsilon_{Nd} = +5.7$ to $-30$) and Sr ($^{87}Sr/^{86}Sr = 0.704462$ to 0.740102) isotopic values (Johnson, 1991; Perry et al., 1993; Wendlandt et al., 1993). Published NCI values for the North American ignimbrites (Perry et al., 1993) show a wide range (0.05 to 0.92), although the systems with a similar tectonic setting to the Neogene Central Andes (i.e., subduction related — San Juan, Grizzly Peak and Kalamazoo systems) have overlapping values (0.43 to 0.92) with the Cerro Galán and APVC systems (0.49 to 0.83; Schmitt, 1999).

Thus, differences between the radiogenic isotopic character of North American silicic systems and those in the Central Andes can be broadly resolved. The large variations in Nd, Pb and Sr isotopic values between regions of silicic ignimbrites in North America may be explained by differences in the thickness, age and type of underlying crust. In contrast, the silicic ignimbrites erupted from the Central Andes formed in a broadly coherent tectonic regime and over consistently thickened crust (up to 70 km; Fig. 4F) in a relatively restricted geographical range (Fig. 1A) and a compressional regime. It is therefore inevitable that magmas residing at shallow levels in North America would interact with a great diversity of crustal lithologies. These factors could therefore explain at least some of the greater radiogenic isotopic variations in the North American ignimbrites compared to ignimbrites in the Central Andes.

### 5.4. Oxygen isotope character of large silicic systems in North America

While differences in radiogenic isotopes will reflect the differences in the age and composition of crust being assimilated and the relative contribution from the mantle (these factors are influenced by tectonic setting), differences in oxygen isotopes may reflect additional processes (Table 3; Bindeman and Valley, 2002). Large volume ignimbrites associated with large silicic systems in North America share the same *intra-unit* oxygen isotope homogeneity seen in

**Table 3**

Physical and chemical characteristics of various large volume ignimbrites in the Central Andes and North America.

| Location | Ignimbrite or sub-unit | Caldera/unit | Volume (km$^3$) | Eruption age (Ma) | T range (°C) | Crystallinity (%) | SiO$_2$ (wt%) | δ$^{18}$O(Quartz) (‰) | δ$^{18}$O(Other) | δ$^{18}$O(Melt)[a] (‰) | References[b] |
|---|---|---|---|---|---|---|---|---|---|---|---|
| *CVZ ignimbrites (this study)* | | | | | | | | | | | |
| Cerro Galan | Blanco | Cerro Galan | 8 | N.D. | 790–820 | 51 | 68–71 | 9.19 ± 0.07 | | 8.62 ± 0.07 | This study, 1,2 |
| | L. Merihuaca | Cerro Galan | 8 | 5.60 ± 0.20 | 790–820 | 52 | 68–71 | 9.53 ± 0.06 | | 9.03 ± 0.06 | |
| | M. Merihuaca | Cerro Galan | 35 | 5.56 ± 0.10 | 790–820 | 50 | 68–71 | 8.84 ± 0.03 | 7.77, Plag | 8.34 ± 0.03 | |
| | U. Merihuaca | Cerro Galan | 20 | 5.49 ± 0.11 | 790–820 | 47 | 68–71 | 9.16 ± 0.02 | 8.25, Plag | 8.66 ± 0.02 | |
| | Pitas | Cerro Galan | 190 | 4.84 ± 0.04 | 790–820 | 46 | 68–71 | 8.94 ± 0.11 | 6.65, Zircon | 8.44 ± 0.11 | |
| | Real Grande | Cerro Galan | 390 | 4.68 ± 0.07 | 790–820 | 45 | 68–71 | 9.10 ± 0.11 | 7.67, Zircon | 8.60 ± 0.11 | |
| | Cueva Negra | Cerro Galan | 50 | 3.77 ± 0.08 | 790–800 | 45 | 68–71 | 9.47 ± 0.10 | | 8.97 ± 0.10 | |
| | Pre-CGI igs | Cerro Galan | N.D.[c] | 2.80 ± 0.04 | 790–800 | N.D. | 68–71 | 9.49 ± 0.02 | | 8.99 ± 0.02 | |
| | CGI | Cerro Galan | 630 | 2.08 ± 0.02 | <790 | 46 | 68–71 | 9.25 ± 0.19 | 6.98, Zircon | 8.75 ± 0.19 | |
| APVC | Artola | Buried | >100 | 9.40 ± 0.03 | 750–800 | 29–38 | 71 | 8.37 ± 0.17 | | 7.87 ± 0.17 | This study, 3,4,5,6 |
| | Guacha | Cerro Guacha | 1300 | 5.65 ± 0.01 | N.D. | N.D. | 67–71 | 8.32 ± 0.03 | | 7.82 ± 0.03 | |
| | Puripicar | Unknown | >500 | 4.09 ± 0.02 | 750–800 | 52–58 | 68 | 9.58 ± 0.17 | | 9.08 ± 0.17 | |
| | Toconao | La Pacana | 300 | 4.5–4.0 | 730–750 | 1–15 | 77 | | 7.71, Plag | 8.11 | |
| | Atana | La Pacana | 2200 | 3.96 ± 0.02 | 770–790 | 30–40 | 66–70 | 8.33 ± 0.04 | | 7.83 ± 0.04 | |
| | Tara | Cerro Guacha | 800 | 3.49 ± 0.01 | N.D. | 25–30 | 67–72 | 8.48 ± 0.12 | 7.66, Plag | 7.99 ± 0.12 | |
| | Purico | Purico | 100 | 1.3–1.0 | 750–810 | 38–59 | 65–66 | 8.47 ± 0.08 | 7.51, Plag<br>7.78, Zircon | 7.97 ± 0.08 | |
| *Basaltic to Dacitic Lavas* | | | | | | | | | | | |
| Barroso Volcanics | | | | 5.5 to recent | | | 55.7–60 | | 6.9–7.9, Bulk Rock | | 7 |
| CVZ (24 to 26° S) | | | | <2 | | | 52.6–56 | | 7.5–8.5, Bulk Rock | | 8 |
| Volcan Ollague | | | | <5.5 | | | 52.9–67 | | 7.1–8.1, Plag | | 9 |
| *North American ignimbrites* | | | | | | | | | | | |
| Oligocene Great Basin Tuffs (GBT) | | | | | | | | | | | |
| SW Nevada | Copper Summit | | N.D. | 35.7 | | <40 | 68–70 | 9.6 to 10.1 | | 9.1 to 9.6 | 10,11 |
| | North Creek | | 50 | 35 | | >20 | N.D. | 9.5 | 7.4 to 8.3, Plag | 7.8 to 8.7 | |
| | Kalamazoo | | 500 | 34.7 ± 0.7 | | >25 | 68–72 | | 7.5 to 8.2, Plag<br>8.1 to 8.4, Sanidine | 7.9 to 8.6<br>8.5 to 8.9 | |
| Central Nevada | Central Nevada & Indian Peak Caldera Complex | | >5000 | 31.7–25.4 | | <55 | 66–78 | 8.8 to 10.5 | 7 to 9, Plag | 8.3 to 10 | 12,13,14 |

| | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Utah/Nevada | Lund Tuff (LT) | White Rock | 3000 | 29.02 ± 0.04 | 770 | 48–56 | 66–67 | 7.5–9.0 | | 7 to 8.5 | 15 |
| San Juan caldera complex | La Jara Canyon | Platoro | 1000 | 29.3 ± 0.3 | | 30–40 | 64–67 | | 6.7 to 6.92, Plag | 7.1 to 7.32 | 16,17 |
| | Ojito Creek | | 100 | 29.1 ± 0.3 | | 25–35 | 68–70 | | 6.82, Plag | 7.22 | |
| | Middle Tuff | | 100 | 29 | | 25–35 | 64–67 | | 5.5 to 6.8, Plag | 5.9 to 7.2 | |
| | Masonic Park Tuff | | 500 | 28.6 ± 0.23 | | 25–35 | 62–68 | | 6.86 to 6.88, Plag | 7.26 to 7.28 | |
| | Fish Canyon Tuff (FCT) | La Garita | 5000 | 28.03 ± 0.18 | | | | | | | 18,19 |
| | Late | | | | 763 | 35–51 | 66–68 | 8.36 ± 0.01 | 4.34, Sphene | 7.82 ± 0.15 | |
| | Early | | | | 762 | 35–51 | 66–68 | 8.13 ± 0.06 | 4.19, Sphene | 7.60 ± 0.15 | |
| | Pre-caldera | Pagosa Peak | | 29.93 ± 0.18 | 751 | 45 | 66–68 | 8.18 ± 0.10 | 4.67, Sphene | 7.63 ± 0.10 | |
| | Post-caldera | Nutras Creek | | 28.06 ± 0.16 | 701 | 44 | 66–68 | 8.41 ± 0.10 | | 7.83 ± 0.10 | |
| | Carpenter Ridge Tuff (CRT) | Bachelor | >1000 | 27.55 ± 0.05 | | 10 | 68–74 | | 7.1 to 7.2, Fspar | 7.5 to 7.6 | 20,21 |
| | Nelson Mountain Tuff | Nelson–Cochetopa | >500 | 26.90 ± 0.02 | | 30–40 | 63–74 | 8.1 to 8.2 | 6.7 to 6.9, Fspar | 7.6 to 7.7 | 20,21 |
| | Snowshoe Mountain Tuff | Creede | >500 | 26.87 ± 0.02 | | 35–45 | 62–66 | 8.1 to 8.2 | 6.3 to 6.7, Fspar | 7.6 to 7.7 | 20,21 |
| SW Nevada Volcanic Field | Topopah Spring Tuff | Timber Mountain/ | 1200 | 12.8 | 800–900 | 10 | 67.3 | | | 8.06 | 22 |
| | Tiva Canyon Tuff | Oasis Valley | 1000 | 12.7 | 800–900 | 11 | 67 | | | 8.13 | |
| | Rainier Mesa Tuff | (TM/OV) | 1200 | 11.6 | 700–800 | 23 | 76.8 | | | 7.25 | |
| | Ammonia Tanks Tuff (ATT) | | 900 | 11.45 | 700–800 | 14–19 | 73.1–75.7 | | | 5.39 to 5.57 | |
| Snake River Plain Tuffs (SRPT) | Cougar Pt Tuff | Bruneau–Jarbridge | >1000 | 12.64–10.8 | 800–1000 | <20 | 72–76 | 2.6 to 4.7 | −1.3 to 3.8, Fspar | 0.1 to 3.9 | 23,24 |
| | Various | Twin Falls | ? | 11.3–9 | 900–1000 | 3–15 | 69–77 | | 1.7 to 3, Plag | | 25 |
| | Blacktail Creek Tuff | Heise | 1200 | 6.62 ± 0.03 | 800–900 | 20 | 74.4 | 6.4 | 4.81, Zircon | 6.6 | 26 |
| | Walcott Tuff | Heise | 750 | 6.27 ± 0.04 | 800–900 | 1–2 | 75.5 | | 4.17, Zircon | 6.1 | 26 |
| | Conant Creek Tuff | Heise | 300 | 5.51 ± 0.13 | 800–900 | 5–7 | 74.3 | | 3.96, Zircon | 5.9 | 26 |
| | Kilgore Tuff | Heise | 1800 | 4.43–4.52 | 800–900 | 1–10 | 75.8–76.4 | 4.29 to 4.33 | −1.3 to 6, Zircon | 3.3 to 3.6 | 26 |
| Central Oregon | Rattlesnake Tuff (RST) | High Lava Plains | 280 | 7.1 | 800–880 | 1 | 75.6–77.7 | 7.6 to 7.8 | | 7.1 to 7.3 | 27 |
| Yellowstone | Huckleberry Ridge Tuff (HRT) | Yellowstone | 2500 | 2.053 ± 0.096 | 850–900 | 2–15 | 74 – 77 | 7.0 to 7.5 | | 6.5 to 7.0 | 28 |
| New Mexico | Lower Bandelier Tuff (LBT) | Toledo | 600 | 1.61 ± 0.01 | | | | | | | 19,29,30 |
| | Late | | | | 771 | 25 | 75 | 7.81 ± 0.10 | | 7.28 ± 0.10 | |
| | Early | | | | 737 | 20 | 77 | 7.90 ± 0.10 | | 7.32 ± 0.10 | |
| Eastern California | Bishop Tuff (BT) | Long Valley | 650 | 0.76 ± 0.002 | | 5–25 | 74 | | | | 19,31,32 |
| | Late | | | | 819 | | | 8.21 ± 0.10 | 5.82, Zircon | 7.80 ± 0.10 | |
| | Early | | | | 743 | | | 8.33 ± 0.10 | 5.61, Zircon | 7.80 ± 0.10 | |
| Yellowstone | Lava Creek Tuff (LCT) | Yellowstone | 1000 | 0.640 ± 0.002 | | | | | | | 19,28 |
| | Late | Member B | | | 912 | 15–25 | 73 | 6.40 ± 0.05 | 5.49, Sanidine | 5.98 ± 0.05 | |
| | Early | Member A | | | 800–850 | 5–20 | 77 | 6.46 ± 0.10 | 5.34, Sanidine | 5.98 ± 0.10 | |

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[a] $\delta^{18}O_{(Melt)}$ values are 0.5 lower than $\delta^{18}O_{(Quartz)}$ values and 0.4 higher than $\delta^{18}O_{(Plag)}$ values; based on mineral-melt fractionation from Matthews et al. (1994) and Bindeman and Valley (2002).

[b] See Supplementary material for reference details.

[c] Not determined.

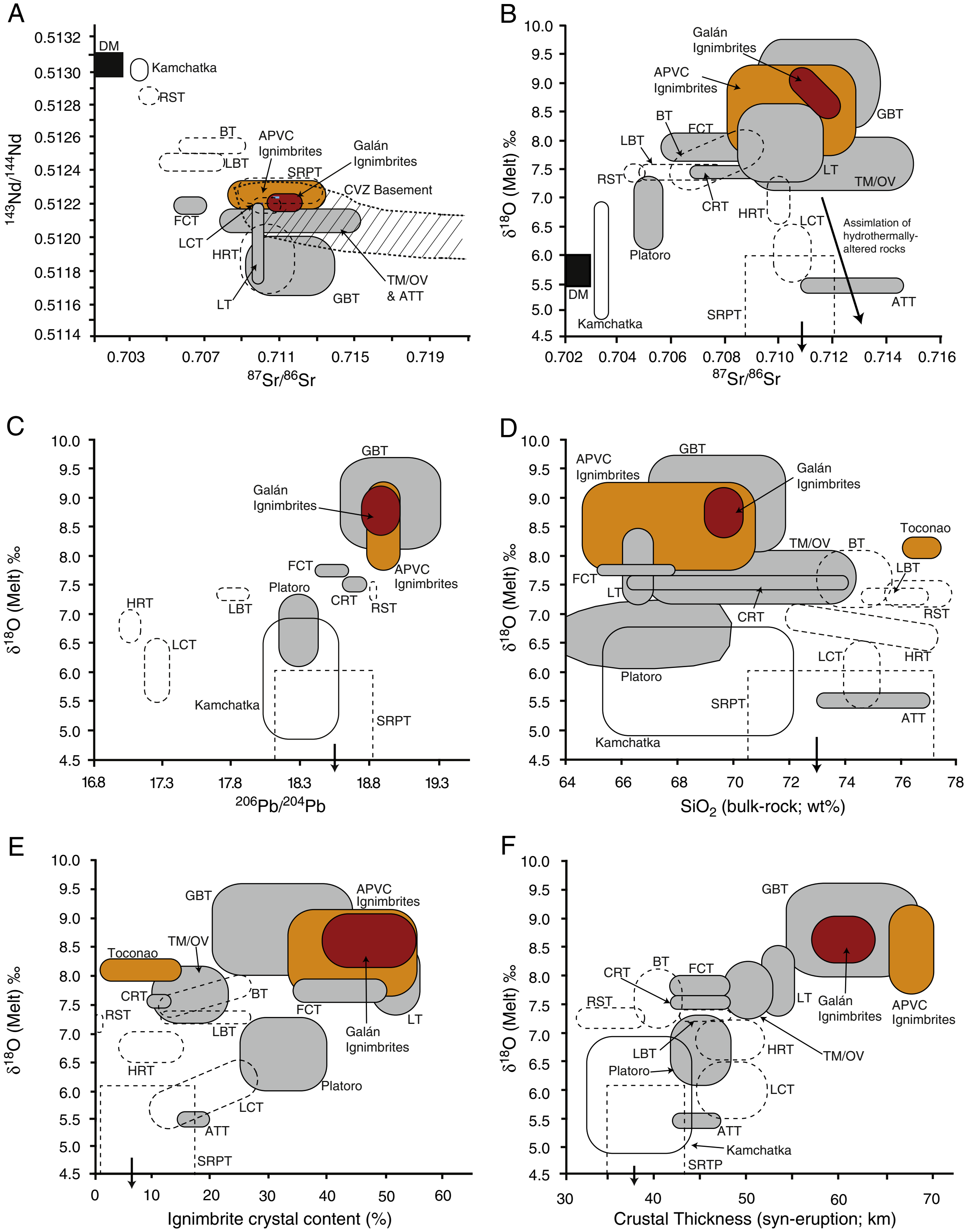

**Fig. 4.** Chemical and physical characteristics of Central Andean and North American ignimbrite magmas. Solid (filled) fields are subduction related or influenced systems in North America and the Central Andes; other systems (dashed areas) are shown for comparison purposes only. Abbreviations for North American ignimbrites: LT = Lund Tuff; FCT = Fish Canyon Tuff; CRT = Carpenter Ridge Tuff; ATT = Ammonia Tanks Tuff; GBT = Oligocene Great Basin Tuffs; HRT = Huckleberry Ridge Tuff; LCT = Lava Creek Tuff; LBT = Lower Bandelier Tuff; RST = Rattlesnake Tuff; BT = Bishop Tuff; SRPT = Snake River Plain Tuffs; TM/OV = Timber Mountain/Oasis Valley; ATT = Ammonia Tanks Tuff; Depleted Mantle composition (DM; Saunders et al., 1988). Fields for Kamchatka ignimbrites (unbroken outline) are also shown for comparison (Bindeman et al., 2010). See Table 3 and Supplementary data for references and details on all ignimbrites in the Central Andes and North America and basaltic andesite lavas in the Central Andes. Current crustal thicknesses from Wigger et al. (1994) and Das and Nolet (1998); Mid-Tertiary crustal thicknesses for the Lund and Great Basin Tuffs, and Tuffs from the San Juan caldera complex from Best et al. (2009).

individual Central Andean ignimbrites. The 'early' and 'late' (or initial and final) portions of many large-volume tuffs in North America show very little variation in $\delta^{18}O$ values, suggesting that the pre-eruption magma chambers were not zoned in terms of oxygen isotopes. For example, the $\delta^{18}O_{(Qtz)}$ values of early and late Bishop Tuff (BT) deposits are +8.33 ± 0.10 and +8.21 ± 0.10‰ respectively, and +6.46 ± 0.10 and +6.40 ± 0.05‰ for early and late deposits of the Lava Creek Tuff (LCT; Table 3). In both North America and

Central Andean large-volume ignimbrite provinces, magmatic systems develop $\delta^{18}O$ homogeneity suggesting that the timescales of magma accumulation must be at least $10^4$–$10^5$ years in order for convection to effectively homogenize these large volume magmas (e.g., Bindeman and Valley, 2002). This agrees with timescales inferred for magma residence of large volume silicic magmas in shallow chambers in both these regions as deduced from U–Pb ages of zircon autocrysts (e.g., Schmitt et al., 2002; Bachmann et al., 2007; Folkes et al., 2011a). In a few systems, however, such as the TM/OV system (Bindeman et al., 2006), there are significant variations in $\delta^{18}O$ (and Sr–Nd) values suggesting syn-eruptive mixing of diverse magma batches.

However, it is important to note that despite the general intra-system $\delta^{18}O_{(Qtz)}$ homogeneity of large-volume magmas in both the Central Andes and North America, there are some important variations in $\delta^{18}O$ values between the two regions. The large volume ignimbrites from North America show a generally lower and wider range in $\delta^{18}O_{(melt)}$ values (+0.1 to +9.6‰), compared to similar systems in the central Andes (+7.8 to +9.1‰; Fig. 4A-F; Table 3) over a similar range in $SiO_2$. Much of this variation is likely attributed to the different tectonic settings in which these magmas were generated and evolved and indeed most of the very low $\delta^{18}O$ magmas come from the anorogenic systems of Yellowstone–Snake River Plain.

To eliminate tectonic setting as a variable, below we focus on the subduction-related or influenced North American systems that are the most comparable with the Central Andes. As described above, these are the Oligocene Great Basin of Nevada and Utah, the SRMVF in Colorado and New Mexico, and the TM/OV caldera systems in SW Nevada. We have shown that the variations in radiogenic isotopic (Sr, Nd and Pb) values between large volume ignimbrites in the Central Andes and North America can be attributed to variations in the age, type and composition of the continental crust hosting each large silicic system. However, we cannot explain the variations in $\delta^{18}O$ values between the two regions in the same manner. The $\delta^{18}O$ values of basement rocks and plutons that are the likely source for mid- to upper-crustal assimilation in large volume magmas from these regions all contain elevated $\delta^{18}O_{(Qtz)}$ values that overlap and exceed those $\delta^{18}O_{(Qtz)}$ values in magmas calculated in this study and in the rest of the scientific literature. For example, Precambrian gneisses in the Oligocene eastern Great Basin region range from +8 to +14‰ (King et al., 2004). The crust in the SRMVF of Colorado (comprising the Fish Canyon, Carpenter Ridge, and Platoro caldera Tuffs) consists of various granites and metasedimentary rocks with $\delta^{18}O$ values between +9 and +16‰ (Johnson and Fridrich, 1990; Johnson, 1991). These are similar to $\delta^{18}O$ values calculated for basement rocks in the central Andes (+9.4 to +16‰; Ort et al., 1996; Kay et al., 2011 and references therein). The $\delta^{18}O$ character of the SRMVF tuffs (Fish Canyon, Carpenter Ridge, and Platoro caldera tuffs) is particularly interesting. They all share a similar Sr, Nd, and Pb isotopic character, but lower $\delta^{18}O_{(melt)}$ tuffs (+7.32 to +5.9‰) are found at the Platoro caldera (Table 3), and a trend leading to a ~1‰ lower $\delta^{18}O$ value is described in the main SRMVF sequence (Larson and Taylor, 1986). At the TM/OV center, the youngest tuff, the Ammonia Tanks member, exhibits low-$\delta^{18}O_{(melt)}$ values (+5.39 to +5.57‰), with a similar crustal radiogenic isotope character as earlier tuffs that have high-$\delta^{18}O$ values (the Topopah Springs, +8.06‰; Tiva Canyon, +8.13‰; and Rainer Mesa, +7.25‰ members; Table 3).

### 5.5. Lack of low-$\delta^{18}O$ magmas in the Central Andes

In the Central Andes, no low-$\delta^{18}O$ characteristics have so far been reported from Neogene large volume silicic magmas. This extends across the compositional spectrum from crystal-rich dacite to crystal-poor high-$SiO_2$ rhyolite. For instance, the crystal-poor Toconao ignimbrite is the only ignimbrite of any significant volume (~300 $km^3$) with a similar composition to the high-Si rhyolitic ignimbrites in North America. However, it has $\delta^{18}O$ values no different from the high-crystallinity, lower $SiO_2$ dacites in the Central Andes (Fig. 4D,E). Furthermore, there is no evidence of progressive lowering of $\delta^{18}O$ values during the development of individual confocal systems. Successive eruptions at Cerro Galán, Cerro Guacha, or La Pacana do not contain low-$\delta^{18}O$ values as would be expected if a similar model (a progression to lower-$\delta^{18}O$ magmas) to the TM/OV and Platoro systems in North America is applied.

The prevailing paradigm for the production of low-$\delta^{18}O$ magmas in North American systems is the large-scale melting or 'cannibalization' of hydrothermally-altered low-$\delta^{18}O$ shallow crustal rocks, formed from interaction with high temperature meteoric fluids (e.g., Hildreth et al., 1991; Bindeman et al., 2006; Watts et al., 2011). In some cases, these magmas are thought to be produced by the rapid remelting of this shallow-level low-$\delta^{18}O$ crust that produces the crystal-poor high $SiO_2$ compositions, often facilitated by successive caldera collapse events that bring 'fresh', near surface hydrothermally altered crust down into the melting zone (e.g., Bindeman et al., 2006). Simakin and Bindeman (2012) demonstrate this with a numerical treatment of these melting processes. This process may account for the late low-$\delta^{18}O$ (+5.39 to 5.57‰) signature of the Ammonia Tanks member (73 to 76 wt.% $SiO_2$) and the low-$\delta^{18}O$ (+5.9 to 7.32‰) crystal-rich dacitic ignimbrites (62 to 70 wt.% $SiO_2$) associated with the Platoro caldera complex (Table 3; Fig. 4D,E). This relatively low-$\delta^{18}O$ signature is thus not restricted to high-$SiO_2$ rhyolites, but to silicic magmas in general that assimilated significant volumes of shallow-level low-$\delta^{18}O$ crust (e.g., Hildreth et al., 1991; Bindeman et al., 2006; Watts et al., 2011).

It is reasonable to expect that a similar process should occur in the Central Andes, but it is not recorded in large ignimbrites. We note that Feeley and Sharp (1995) suggest that the 'low' $\delta^{18}O$ values (+7.1 to 8.1‰) in dacite lava flows from Volcan Ollagüe require assimilation of low-$\delta^{18}O$ hydrothermally altered wall rocks during differentiation in shallow crustal magma chambers. However, this concerns very small volume lava flows that are more mafic and with a different magmagenesis to the large volume silicic ignimbrites (see de Silva, 2008) we discuss in this study. So while local interactions may be recorded, there is no evidence for large-scale interaction of the silicic magmas with hydrothermally altered rock.

Following the logic that availability of meteoric water will govern the health and vitality of hydrothermal systems (e.g., Hedenquist et al., 1998; So et al., 2005), the lack of a low-$\delta^{18}O$ magmatic signal might indicate a paucity of low-$\delta^{18}O$ meteoric water in the Central Andes. If large volumes of low-$\delta^{18}O$ meteoric water were available (either via surface water, groundwater or precipitation), it should percolate into the shallow crust and fuel extensive hydrothermal systems above the Neogene silicic magmatic systems. Given that the regional permeability of near surface deposits in North America and the Central Andes is similar ($\log k = -12$ to $-17$ $m^2$ – Central Andes; $\log k = -11$ to $-16$ $m^2$ – western North America; Gleeson et al., 2011), regional permeability variations are not a contributing factor to the disparate $\delta^{18}O$ data in these two magmatic provinces. The fact that we do not see a low-$\delta^{18}O$ signature in any of the large-volume Central Andean magmas may thus indicate a limited supply of low-$\delta^{18}O$ meteoric water present in this elevated plateau since at least 10 Ma. Alternatively, vigorous hydrothermal systems may have been available but were never ingested into the magmatic systems. We note that surface hydrothermal activity in the Central Andes is limited. The El Tatio and Sol de Mañana geothermal zones are the only regions of note (de Silva and Francis, 1991), although small thermal springs are found in almost all the major caldera systems in this region. In this context, we find the evidence for low precipitation and high aridity in the Central Andes throughout the last 10 Ma (Strecker et al., 2007; Barnes and Ehlers, 2009; Bywater-Reyes et al., 2010), coincident with the period of large volume silicic volcanism we deal with here, intriguing with respect to the lack of a low-$\delta^{18}O$

signature. Below, we explore the possibility that the prevailing climate may be a factor in determining the O-isotope record of these large silicic volcanic systems. Supporting this is the O-isotope record of such systems in Kamchatka.

### *5.6. Oxygen isotope character of large silicic systems in Kamchatka*

The Kamchatkan Peninsula in eastern Russia is built on continental crust and is part of the Kurile–Kamchatka volcanic arc related to the subduction of the Pacific Plate with volcanism spanning significant ranges in age, volume, and $\delta^{18}$O values (Bindeman et al., 2004; Bindeman et al., 2010 and references therein). Thus, Kamchatka represents a 'microcontinent' type volcanic arc environment and a good and proven end-member case to test the proposal that climate may exert an important control on the $\delta^{18}$O values of crustal magma systems. It is important to note that Kamchatka, despite being built on high-$\delta^{18}$O basement rocks and being influenced by high-$\delta^{18}$O mantle sources, has the largest concentration of low-$\delta^{18}$O magmas in the world (e.g. Bindeman et al., 2004). If counted by number of units, more than 50% of all large volume ignimbrites (>10 km$^3$) plot in the low-$\delta^{18}$O magma field (<5.5‰, Bindeman et al., 2010), and display 3–5 *per mil* depletion relative to the mantle sources. Using $\varepsilon_{Nd}$ values for the silicic magmas, the melted crust and basaltic parent magmas (Bindeman et al., 2004), NCI values of 0.1 to 0.4 are obtained. Although this is lower than values for the Central Andes and the subduction-related systems in North America (because of the less radiogenic Nd silicic magmas), the Kamchatkan ignimbrite magmas still require the assimilation of many 10's of % of hydrothermally-altered crustal material. We consider Kamchatka as a proven case of $\delta^{18}$O-depletion because of documented widespread assimilation of hydrothermally-altered rocks, altered by synglacial meltwaters during the Pleistocene glaciation. We observe that Kamchatka plots on the lower end of $\delta^{18}$O values on all our graphs (Fig. 4), anchoring the overall trend from S America (high) to N America (very diverse) to Kamchatka (universally low) $\delta^{18}$O values. These low values correlate with a decrease in average crustal thickness and elevation (Figs. 4F, and 5), an increase in the amount of precipitation (Fig. 6A), and interaction with extremely light (<−25‰) glacial meltwaters (Chesko, 1994; Bindeman et al., 2004). We note here that the regional permeability in Kamchatka (log k = −12 to −15 m$^2$; Gleeson et al., 2011) is similar to values in North America and the Central Andes.

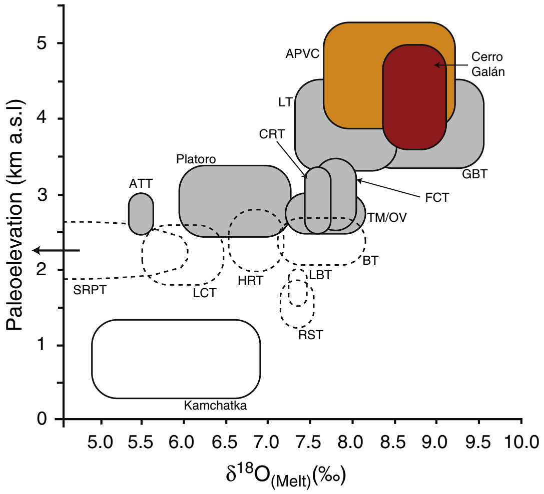

**Fig. 5.** Oxygen isotopic composition vs. paleoelevation for large volume silicic systems in North America and the Central Andes. Solid (filled) fields are subduction related or influenced systems in North America and the Central Andes; other systems (dashed areas) are shown for comparison purposes only. Abbreviations as per Fig. 4 caption. Paleoelevation data from the following sources; Central Andes — (Strecker et al., 2007 and references therein); North America — (Chase et al., 1998; Link et al., 1999; DeCelles, 2004; Sheldon and Retallack, 2004; Mulch et al., 2008; Best et al., 2009; Busby and Putirka, 2009; Roy et al., 2009; Molner, 2010); and Kamchatka — (Chesko, 1994; Bindeman et al., 2010).

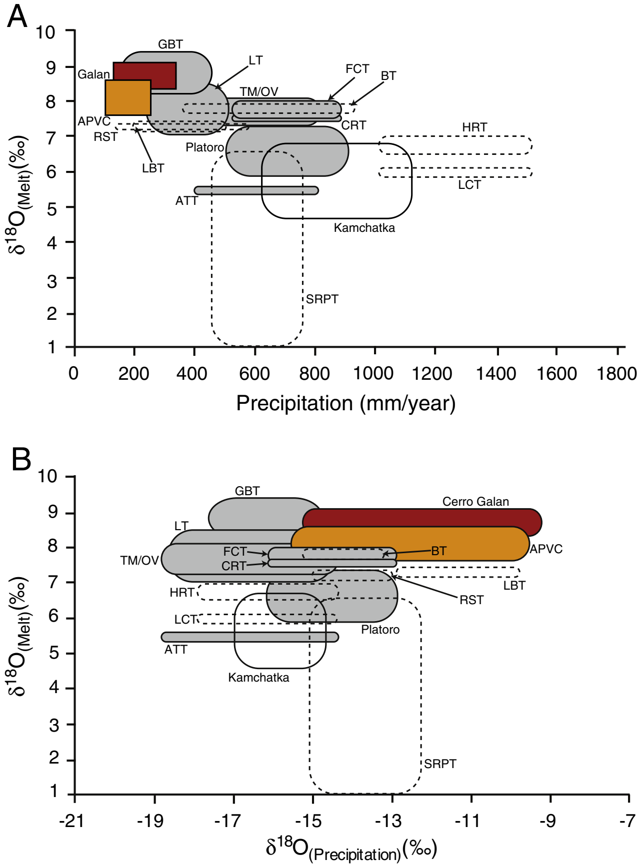

**Fig. 6.** Oxygen isotopic composition of magmas vs. (A) the amount of precipitation and (B) the $\delta^{18}$O of precipitation during the eruption of various large volume silicic ignimbrites in the central Andes and North America (1 sigma uncertainties are shown). Solid (filled) fields are subduction related or influenced systems in North America and the Central Andes; other systems (dashed areas) are shown for comparison purposes only. Abbreviations as per Fig. 4 caption. Data from the following sources; Central Andes (Strecker et al., 2007); North America (Chase et al., 1998; Kendall and Coplen, 2001; Mulch et al., 2008); and Kamchatka (Chesko, 1994; Bindeman et al., 2010). Modern-day $\delta^{18}$O(precipitation) values from Bowen (2012) — see text for discussion.

### *5.7. Climate recorded in the oxygen isotopic composition of Central Andean, Western North American, and Kamchatkan ignimbrites*

Although there have been variations in elevation and aridity in the Central Andes in the Late Cenozoic, for at least the last 10 Ma the region has been an elevated (generally >3500 m; Fig. 5), arid plateau with very low precipitation and high evaporation rates, steep, relatively narrow drainage basins and a corresponding large depth to water tables (Strecker et al., 2007; Barnes and Ehlers, 2009; Bywater-Reyes et al., 2010). The current annual precipitation in the central Andes (100 to 350 mm/year; Fig. 6A) is thought to have been approximately constant over the last 10 Ma (Strecker et al., 2007). Combined, these factors could conspire to limit large amounts of low-$\delta^{18}$O meteoric surface and groundwater permeating down into shallow to intermediate crustal depths and from producing the widespread hydrothermal alteration of shallow crustal material needed to produce low-$\delta^{18}$O magmas. Assuming the current $\delta^{18}$O range for precipitation in the Central Andean region (−9 to −15‰; Fig. 6B; Bowen, 2012) was approximately the same during the Late Miocene to Pliocene (allowing for fluctuations within this range), the high-$\delta^{18}$O of the magmas argues against substantial interaction with precipitation/meteoric water via extensive hydrothermal systems. Indeed, Quade et al. (2007) showed that for every km of elevation, the $\delta^{18}O$ value of precipitation falls by ~2.8‰. Our reasoning leads to the conclusion that large-scale assimilation of highly $\delta^{18}O$-depleted hydrothermally altered crust is not a factor during the production, accumulation and storage of large volume silicic magmas in the Central Andes because very low precipitation and high evaporation rates, resulting from the high elevations (thick crust) limit extensive hydrothermal systems from developing.

Constraining paleoelevation and paleoprecipitation of the caldera systems in western North America is more difficult due to the large spatial and temporal range covered by these systems. Modern-day data is less meaningful for these older systems such as those in the Oligocene Great Basin region. However, we have attempted to offset this uncertainty by using the widest range of $\delta^{18}O$ values for modern-day meteoric waters (Bowen, 2012). Clearly, any hydrothermal alteration of near-surface deposits would dramatically lower the $\delta^{18}O$ values of crust that is assimilated into these magmatic systems. Data from a range of studies has been sourced and is shown in Fig. 5 (paleoelevation) and Fig. 6 (paleoprecipitation). We have focused our discussion on the subduction-related or influenced systems in North America but use various others (such as the Snake River Plain–Yellowstone) for comparison. Chase et al. (1998) provide a useful overview and starting point, but region-specific data comes from the following sources; the Oligocene Great Basin of Nevada and Utah (DeCelles, 2004; Sheldon and Retallack, 2004; Best et al., 2009), the Sierra Nevada (Mulch et al., 2008; Busby and Putirka, 2009; Molner, 2010), the Snake River Plain and Yellowstone region (Link et al., 1999), and the Colorado Plateau (Roy et al., 2009).

The most directly analogous North American examples to the central Andean magmas are the large-volume ignimbrites erupted during the Mid-Tertiary from the subduction-related systems of the eastern Great Basin in Nevada and Utah. Interestingly, $\delta^{18}O_{(Qtz)}$ values of magmas for this region are the highest of all samples in North America (+9.5 to +10.1‰). Applying our reasoning from the Central Andes, this geochemical signature could be due to a drier climate as a consequence of thickened crust and associated high elevations prior to and during the eruption of these Great Basin Tuffs. Indeed, recent work posits that the Oligocene ignimbrites (the eastern Great Basin Tuffs including the Lund Tuff) in this region were erupted onto a flat plateau with a similar elevation (up to 4500 m; Fig. 5), crustal thickness (up to 70 km; Fig. 4F) and aridity (100 to 500 mm/year of precipitation; Fig. 6B) to the Neogene Central Andean Altiplano-Puna. Due to these similarities in the climate and physiography of the Central Andes over the past 10 Ma and Great Basin region prior to and during the eruption of the Great Basin and Lund Tuffs (~36 to 29 Ma), this region has been termed the 'Nevadaplano' (e.g., DeCelles, 2004; Best et al., 2009). The western margin of this high plateau has been proposed to have acted as a topographic barrier to ignimbrite flow and also as a heavy orographic precipitation barrier creating arid conditions in the Nevadaplano, inhibiting meteoric water production of hydrothermal systems that would have imparted a depleted $\delta^{18}O$ signature to the magmas that subsequently erupted large volume silicic ignimbrites. The presence of westward-draining stream valleys would have helped to channel low-$\delta^{18}O$ water away from the upper crust in the interior of this plateau (Best et al., 2009). This (along with thicker crust) may help to explain the higher $\delta^{18}O$ values of large volume silicic magmas in this region compared with other systems in the western US (Fig. 4).

Some systems in the Oligocene Great Basin show evidence of surface and shallow crustal interaction with low-$\delta^{18}O$ meteoric waters. John and Pickthorn (1996) examined the hydrothermally-altered intracaldera tuff and shallow plutons of the Stillwater caldera complex (~29 to 28 Ma) in the Great Basin of western Nevada and found $\delta^{18}O$ values as low as −6 to −9‰. Intracaldera tuffs and intrusions of the Indian Peak caldera are also extensively propylitically altered from interaction with meteoric water (E. Christiansen, pers. comm., April 2013). Multiple calderas appear to show progressive lowering of $\delta^{18}O$ values in magmas, especially when there is significant nested overlap (our unpublished data). Although there is evidence for the presence of meteoric water and hydrothermally altered shallow crustal materials in this region, this is not reflected in the $\delta^{18}O$ signature of the large volume Oligocene Great Basin magmas as a whole. Perhaps the shallow depth of this altered material and/or the relatively small proportions that might be assimilated into the large volume magmas may be limited by the high paleoelevation and low paleoprecipitation in this region.

Grunder and Wickham (1991) identified a general $\delta^{18}O_{(melt)}$-depletion over time in the east-central Nevada Great Basin, from +8.9 ± 0.2‰ in the Early Group to +7.9 ± 0.3‰ in the Late Group Tuffs (~38.5 to 34 Ma). These authors attribute this depletion to a decrease in the $\delta^{18}O$ values of the crustal rocks that undergo assimilation to produce the large volume magmas in this region. These workers argued that this was achieved by a progressive increase in the input of deep crustal mantle-derived igneous rocks or fluids derived from them (Grunder and Wickham, 1991). Interaction with shallow low-$\delta^{18}O$ meteoric waters, although recognized as an important localized occurrence, was eschewed as the cause of this widespread $\delta^{18}O$-depletion due to unreasonably high porosities required, or the unreasonable assimilation of large volumes of more realistically lower porosity upper crustal rocks (Grunder and Wickham, 1991). We note that despite this depletion, and the apparent presence of hydrothermally-altered intracaldera volcanic fill and associated shallow intrusions in some calderas, the primary signal for the Great Basin systems is an 'enriched' $\delta^{18}O$ magmatic signature. To us, this affirms that the elevation and aridity of the Nevadaplano region in the Mid-Tertiary helped to reduce the availability of low-$\delta^{18}O$ meteoric waters for interaction with the shallow crust, thereby maintaining the high (albeit lowered over time) $\delta^{18}O$ values of these large volume magmas in the Great Basin region.

Most Kamchatkan ignimbrites have low-$\delta^{18}O$ values and plot as an end member case in Fig. 6. The presently low-$\delta^{18}O$ values in precipitation in Kamchatka (e.g., Chesko, 1994) could have been even lower during the Pleistocene glaciations, as most ignimbrite deposits are intricately intercalated with glacial moraines (Bindeman et al., 2010). Regardless of the geographical position, the Kamchatkan crust, similar to the subglacial crust in Iceland, has been regionally fingerprinted by low-$\delta^{18}O$ values, but depletions primarily occurred in calderas and areas of high heat fluxes that promoted water–rock interaction. The presence of glaciers inside, or in the vicinity of calderas, as is observed in Kamchatka (e.g. Bindeman et al., 2010) creates the best hydrogeologic conditions for isotopic 'conditioning' of the crust and depletion in $\delta^{18}O$ values.

## 6. Conclusions

The Neogene Central Andean Ignimbrite Province is a unique natural laboratory to study the geochemical evolution of large silicic systems recorded in repeated ignimbrite eruptions with different volumes and repose periods. Using $\delta^{18}O$ data, we have shown that this region has produced homogeneous magmas with a close family resemblance over ~10 Myr. The elevated, restricted $\delta^{18}O_{(Qtz)}$ values (up to +9.6‰) and the lack of low-$\delta^{18}O$ values compared with similar magmatic systems in North America and Kamchatka are proposed to strongly reflect the lack of interaction between magmas and extensive shallow hydrothermal systems that, if cannibalized, would have imparted a low-$\delta^{18}O$ signature to the magmas. This could reflect conditions that simply limited interaction between magmas and hydrothermal systems (e.g., that the high-$\delta^{18}O$ magmas are somehow self-sealing). However, given the evidence that the Central Andes has been an elevated, arid plateau for at least the last 10 Ma, we prefer a model where the availability of low-$\delta^{18}O$ meteoric waters that fuel shallow hydrothermal systems was limited. The uniformly high-$\delta^{18}O$ values of Neogene silicic ignimbrite-producing magmas thus not only reflect extensive assimilation of continental crust during silicic magmagenesis, but also signal the high elevation and resulting arid climate of the region.

In contrast, many of the large volume ignimbrites in North America and Kamchatka have lower $\delta^{18}$O values that reflect the influence of wetter climates, lower evaporation rates and lower elevations. These physiographic conditions would have promoted the retention of low-$\delta^{18}$O near-surface meteoric waters that through interaction with magmas, causes a general lowering of magmatic $\delta^{18}$O values. In these systems, large amounts of low-$\delta^{18}$O hydrothermally-altered shallow crustal material were assimilated and/or remelted to produce low $\delta^{18}$O magmas. Ignimbrites of the Mid-Tertiary 'flare-up' of the eastern Great Basin of the western US are the most analogous to those in the Central Andes. They have high-$\delta^{18}$O values and share many chemical, petrological and physical characteristics with those from the Central Andes lending support to their proposed association with a high and dry climate of a putative "Nevadaplano" plateau.

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## Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.jvolgeores.2013.05.014.

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