

## ***Supplementary Material***

### **1 Analytical and Micro-analytical Methods**

#### **Bulk Rock Analyses of the Starting Materials**

Laser fluorination oxygen isotope analyses of starting serpentinite and basalt were performed at the University of Oregon (USA) stable isotope laboratory (Bindeman, 2021) using a MAT 253 mass spectrometer integrated to the laser line using a home-built airlock sample chamber. Three aliquots of University of Oregon Garnet (UOG) reference material ( $\delta^{18}\text{O} = +6.52\text{‰}$ ) were repeatedly analyzed together with the unknown samples during each analytical session and varied no more than  $\pm 0.2\text{‰}$  ( $1\sigma$  standard deviation). The data are represented in Table S1.

#### **Secondary Ion Mass Spectrometry (SIMS)**

Oxygen isotopes in the experimental products were analyzed by secondary ion mass spectrometry (SIMS) with the CAMECA IMS 1270 E7 ion probe at CRPG-CNRS (Nancy, France), where  $^{18}\text{O}/^{16}\text{O}$  ratios were measured simultaneously in the multi-collection mode, using L'2 and H1 off-axis Faraday caps for counting  $^{16}\text{O}^-$  and  $^{18}\text{O}^-$  ions, respectively. The determination of  $\delta^{18}\text{O}_{\text{VSMOW}}$  relative to Vienna Standard Mean Ocean Water for the forsteritic olivines in the P32 and P33 was based on the San Carlos Fo<sub>91.3</sub> ( $\delta^{18}\text{O}_{\text{VSMOW}} = 5.35 \pm 0.09\text{‰}$ ,  $1\sigma$ ) and CI-114 Fo<sub>74</sub> ( $\delta^{18}\text{O}_{\text{VSMOW}} = 5.25 \pm 0.09\text{‰}$ ,  $1\sigma$ ) reference olivines (Bindeman, 2008; Gurenko et al., 2016) analyzed in the bracketing mode (for more details, see Borisova et al., 2016). The data are represented in Table S5.

#### **Laser Ablation Inductively Coupled Plasma Mass Spectrometry (LA-ICP-MS)**

Laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) was applied at the Max Planck Institute for Chemistry, Mainz (Germany). Major and trace element contents were determined by LA-ICP-MS using a New Wave 213 nm Nd:YAG laser UP 213, coupled with a sector-field ICP-MS Element2 (Thermo Scientific) (Jochum et al., 2007). The repeatability (RSD) of the measurements is about 1 - 3% (with measurements corresponding to laser crater of 30  $\mu\text{m}$ ) and 5 - 10% (with measurements corresponding to laser crater of 8  $\mu\text{m}$ ). More details on the acquisition may be found in Borisova et al. (2021). The detection limits for the 30  $\mu\text{m}$  measurements vary between about 0.001 and 1 ppm (Jochum et al., 2014). The concentration values agree within about 5% (30  $\mu\text{m}$ ) to 10% (8  $\mu\text{m}$ ) of the reference values (GeoReM database) (Jochum et al., 2005). The data are represented in Table S2.

### **2 Modelling**

#### **Thermodynamic Modelling of Felsic Liquid Crystallization and Model Zircon Crystallization**

2.1. Thermodynamic modelling was applied to obtain information on mineral and fluid phases appearing during equilibrium crystallization of felsic melts during isobaric cooling at 0.2 GPa and also simulate the widespread evolution of felsic liquids produced in lithospheric settings. This thermodynamic modelling provides insight into complementary equilibrium conditions not accessible from laboratory experiments of limited duration. The modelling was performed using the rhyolite-MELTS (version 1.2.0) software (Gualda et al., 2012) at conditions of the melt saturation with aqueous fluid and oxygen fugacity corresponding to two logarithmic units

above quartz-fayalite-magnetite mineral buffer (QFM+2). Modelling of mineral crystallization of equilibrium and fractional crystallization is presented in Table S3.

2.2. Modelling of zircon crystallization is presented in Table S4 and Fig. 2. The primary thermodynamic condition for zircon nucleation and growth from a felsic melt is supersaturation in zircon during cooling of the melt ( $C_{\text{sat}} \leq C_{\text{melt}}$ ) (Bindeman and Melnik, 2016). Following Boehnke et al. (2013), zircon saturation  $C_{\text{sat}}$  in felsic melts is a function of temperature  $T$  (in Kelvin) and a compositional factor  $M = (\text{Na} + \text{K} + 2\text{Ca})/(\text{Al} \times \text{Si})$  which is calculated from the molar proportions of the elements in the felsic glasses:

$$C_{\text{sat}} = 500000 / \exp\left(\frac{10108}{T} - 1.16(M - 1) - 1.48\right) \quad (\text{eqn. 1}).$$

Values of  $C_{\text{sat}}$  are presented in Table S4 and Fig. 2. For the measured melts initially containing 49 - 110 ppm Zr (49 ppm is minimum, 110 ppm is the maximum Zr contents and 78 ppm is the average Zr content), zircon growth may be initiated at temperatures below 750 °C for all values of  $M$ . We also implemented a linear dependence of  $M$  factor on temperature following Harrison et al. (2007) at each computational step:

$$M(T) = 0.5 + 0.0013 (T - 273.15) \quad (\text{eqn. 2}).$$

The equation 2 calculates  $M(T)$ ,  $M$  factor as a function of  $T$  (in Kelvin).  $C_{\text{sat}}$  (marked as  $M(T)$  on Fig. 2) is calculated according to eqn. 1 and values  $M(T)$  from the eqn. 2. Zircon will start to crystallize at temperature below 750 °C. If the felsic melt contains 78 ppm Zr (the average Zr content in the measured felsic glasses), and the  $M$  factor ranges from 1.4 to 2.9 (in the case of P32, P33, P35, P37, P42 SB7, SBbis2 and SBter3 glasses, Table S1), zircon will start to crystallize at even lower temperatures.

An alternative evaluation for zircon saturation  $C_{\text{sat}}$  in felsic melts can be carried out using equation 3 taken from Watson and Harrison (1983):

$$\ln D_{\text{zrc/melt}} = -3.80 - (0.85(M - 1)) + \frac{12900}{T} \quad (\text{eqn. 3}),$$

where  $D_{\text{zrc/melt}}$  is the equilibrium zircon-melt partition coefficient,  $M$  is  $M$  factor and  $T$  is absolute temperature in Kelvin. Overall, this approach confirms that felsic melts are able to crystallize zircon below 750°C (Table S4).

2.3. Additionally, we calculated required temperature of serpentinite formation due to pre-glacial seawater ( $\delta^{18}\text{O}_{\text{VSMOW}} = -0.75 \text{ ‰}$ ) interaction with peridotite protocrust ( $\delta^{18}\text{O}_{\text{VSMOW}}^{\text{Rock1}}$ )

to generate  $\delta^{18}\text{O}_{\text{VSMOW}}$  values in zircons ( $\delta^{18}\text{O}_{\text{VSMOW}}^{\text{Zrn}}$ ) higher than mantle value. According to the model (Rock1),  $\delta^{18}\text{O}_{\text{VSMOW}}^{\text{Rock1}}$  is obtained following the equations:

$$\delta^{18}\text{O}_{\text{VSMOW}}^{\text{Rock1}} = \Delta_{\text{mineral-H}_2\text{O}} - 0.75 \quad (\text{eqn.4}),$$

where  $\Delta_{\text{mineral-H}_2\text{O}}$  fractionation factor (‰) depending on temperature (°C) and the mineral (serpentine) is after Savin and Lee (1988). According to the mixture between the serpentinite (SERP) with basalt (BAS) (Rock2: 50 wt% SERP and 50 wt% BAS) and (Rock3: 80 wt% SERP to 20 wt% BAS),  $\delta^{18}\text{O}_{\text{VSMOW}}^{\text{Rock2}}$  and  $\delta^{18}\text{O}_{\text{VSMOW}}^{\text{Rock3}}$  are obtained following the equations:

$$\delta^{18}\text{O}_{\text{VSMOW}}^{\text{Rock2}} = 0.5 \times \delta^{18}\text{O}_{\text{VSMOW}}^{\text{Rock1}} + 0.5 \times 5.8 \quad (\text{eqn. 5})$$

$$\delta^{18}\text{O}_{\text{VSMOW}}^{\text{Rock3}} = 0.8 \times \delta^{18}\text{O}_{\text{VSMOW}}^{\text{Rock1}} + 0.2 \times 5.8 \quad (\text{eqn. 6}),$$

where 0.2, 0.5 and 0.8 are fractions of the rock components (hydrous peridotite protolith and basalt) in the mixture and  $\delta^{18}\text{O}_{\text{VSMOW}} = 5.8$  ‰ in the basalt.

To calculate the oxygen isotope composition of the model zircon, olivine-quartz-zircon fractionation factor (Chiba et al., 1989; Valley et al., 2003) was applied (Fig. 3; Table S5). The model zircon oxygen isotope compositions ( $\delta^{18}\text{O}_{\text{VSMOW}}^{\text{Zrn}}$ ) are calculated based on the measured composition of the olivine (Table S5) or the bulk-rock compositions (Table S1):

$$\delta^{18}\text{O}_{\text{VSMOW}}^{\text{Zrn}} = \delta^{18}\text{O}_{\text{VSMOW}}^{\text{Rock2/Rock3}} + \Delta_{\text{Ol-Qtz-Zrn}} \quad (\text{eqn. 7}),$$

where fractionation factor of  $\Delta_{\text{Ol-Qtz-Zrn}}$  is  $\sim 1$  ‰ at 750°C, (Ol – olivine, Qtz – quartz and Zrn – zircon).

2.4. Finally, we performed geochemical modelling of the model zircon compositions (Fig. 4) according to the zircon/melt partition coefficients for Y, Nb, REE, Hf, Th and U according to the approach of Claborn et al. (2018), temperature of 700°C, Ti contents in zircon according to Ferry and Watson (2007) and compositions of the 0.2 GPa felsic melts analyzed by LA-ICP-MS (Table S2).

KREEP model has been performed with the KREEP basalt 15386 (Neal and Kramer, 2006) and the starting basalt composition (Borisova et al., 2021), suggesting 10% of the KREEP component in the mixture with the basalt. The composition of the model zircon has been obtained from the mixture of KREEP with basalt provided the hybrid melt composition reflects that of the parental felsic liquid (Fig. 4).

## Supplementary References

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