S1 MAGMARS algorithm

The first version of MAGMARS is included to this submission. The code will also be hosted at the following address: [add url]. Updates will we added as they become available.

The following equations are solved at decreasing $P$–$T$ conditions (in GPa and °C) along a mantle adiabat (polybaric mode) or at fixed $P$ and increasing $T$ (isobaric mode). In polybaric mode, $P$ is decreased in small increments (0.001 GPa by default) and the $T$ is
adjusted as follow:

\[ T \leftarrow \begin{cases} T - 15\Delta P, & \text{subsolidus,} \\ T - F \frac{L}{C_p}, & \text{with melting,} \end{cases} \quad (S1) \]

with \( F \) the melt fraction, \( L \) the latent heat and \( C_p \) the specific heat capacity.

The concentration of each element \( i \) in the melt (\( C_{l}^i \)) and the temperature of the solidus (\( T_{\text{eff,sol}} \)) are evaluated for \( F \) increments of 0.001 (0.1 wt.%). The concentrations of incompatible elements are calculated using the non-modal batch melting equation (section 1.1), and are then used, along with \( P \) and \( \text{Mg#}_{\text{liq}} \), to calculate the concentrations of major elements and \( T_{\text{eff,sol}} \) (section 1.2). The melt is only removed from the mantle (i.e., the residue) if \( F \) exceeds the critical melt fraction (\( \theta = 0.4\text{–}2.0 \) wt.%). The total melt produced in all melting increments is assumed to be pooled in the crust (section 1.3).

### S1.1 Melt concentration of incompatible elements

Melting reactions:

\[ \sum_i \alpha_i = 1, \quad \sum_i \alpha_i \Delta X_i = \Delta F \quad (S2) \]

Mantle modes:

\[ \sum_i X_i = 1 \quad (S3) \]

\( \alpha_i \) is the proportion in which the \( i \)th phase enters the melt and is a function of \( P \) (Table S1). A negative \( \alpha_i \) indicates that melting is incongruent and that the solid phase is produced during melting. \( X_i \) is the proportion of the \( i \)th phase in the mantle before melting begins and is a function of pressure (Table S2).

| \( P \) (GPa) | opx | cpx | spin | oliv |
|---------------|-----|-----|------|------|
| lherz. \( \leq 1.0 \) | 0.47 | 0.8 | 0.08 | -0.35 |
| lherz. 1.5 | -0.9 | 1.6 | 0.08 | 0.22 |
| lherz. 2.0 | -1.2 | 1.9 | 0.08 | 0.22 |
| lherz. \( \geq 2.5 \) | -1.8 | 2.4 | 0.08 | 0.32 |
| harz. 0.5–5.0 | 0.76 | 0 | 0.06 | 0.18 |

**Table S1:** Melting coefficients \( \alpha_i \) as a function of \( P \) for equation S2 based on the experiments of *Collinet et al.* (2015) and this study. For harzburgite melting, the melting coefficients are assumed to be independent of \( P \). In between two melting reactions (e.g., between 1.5 and 2.0 GPa), the melting coefficients are simply assumed to evolve linearly.

| \( P \) (GPa) | opx | cpx | spin | oliv |
|---------------|-----|-----|------|------|
| lherz. \( \leq 1.0 \) | 0.37 | 0.12 | 0.025 | 0.515 |
| lherz. 1.5 | 0.188 | 0.272 | 0.025 | 0.515 |
| lherz. 2.0 | 0.118 | 0.342 | 0.025 | 0.515 |
| lherz. \( \geq 2.5 \) | 0.076 | 0.384 | 0.025 | 0.515 |

**Table S2:** Mantle mode \( X_i \) as a function of pressure for equation S3. In between these pressures, the mode is assumed to evolved linearly.
Table S3: Regression for partition coefficients (equation S4). The coefficients were regressed from the experiments of Collinet et al. (2015) and this study for Na, Al, Ti and P. All coefficients < 0.01 we simply assigned.

Partition coefficients between each phase $\phi$ and the liquid, for each element $i$:

$$D_{\phi i} = a_{i\phi} + b_{i\phi}P + c_{i\phi}P^2$$  \hspace{1cm} (S4)

The parameters $a_{i\phi}$, $b_{i\phi}$ and $c_{i\phi}$ are detailed in Table S3.

The partition coefficient of the $i$th oxide weighted by the proportions in which each of the $n$ minerals enters the melt is

$$p_i = \sum_{\phi=1}^{n} \alpha_{\phi} D_{\phi i}.$$  \hspace{1cm} (S5)

The bulk partition coefficient of the $i$th oxide for a rock with $n$ different mineral phases is

$$D_{si} = \sum_{\phi=1}^{n} X_{\phi} D_{\phi i}.$$  \hspace{1cm} (S6)

Finally, the concentration of incompatible elements in the melt is obtained with the non-modal batch melting equation:

$$C_{i}^l = \frac{C_{i}^s}{F(1 - p_i) + D_{si}}.$$  \hspace{1cm} (S7)

S1.2 Temperature and major element concentrations of the melt

$$\text{Mg\#}_{\text{liq}} = \frac{\text{Mg\#}_{\text{sol}} K_{D}^{\text{oli-liq}}}{\text{Mg\#}_{\text{sol}}(K_{D}^{\text{oli-liq}} - 1) + 1}.$$  \hspace{1cm} (S8)
with \( K^{\text{cl-liq}}_D = \frac{R_{\text{cl}}/\text{Mg}^{\text{cl}}}{R_{\text{cl}}/\text{Mg}^{\text{liq}}} = 0.36 \) (initial approximation).

Then the major element concentrations can be calculated using linear regression models with \( \text{Mg}^{\#}_{\text{liq}}, C^l_i \) and \( P \) as predictive variables.

Lherzolite melting:

\[
C^l_{\text{SiO}_2} = 61.6(1.6) + 1.4(3)P - 12.5(1.7)P^{0.3} - 11.7(8)\text{Mg}^{\#}_{\text{liq}} + 0.42(3)C^l_{\text{Al}_2\text{O}_3} - 0.50(6)C^l_{\text{Na}_2\text{O}}
\]

\[
+ 0.08(8)C^l_{\text{K}_2\text{O}} - 0.94(9)C^l_{\text{P}_2\text{O}_5} + 0.70(5)\frac{C^l_{\text{Na}_2\text{O}} + C^l_{\text{K}_2\text{O}}}{P} - 0.15(5)C^l_{\text{P}_2\text{O}_5}P
\]  

(S9)

\[
C^l_{\text{CaO}} = -0.6(6) - 3.4(2)P + 10.4(1.0)P^{0.3} + 9.6(5)\text{Mg}^{\#}_{\text{liq}} - 0.051(2)C^l_{\text{Al}_2\text{O}_3} - 0.38(3)C^l_{\text{Na}_2\text{O}}
\]

\[
- 0.09(6)C^l_{\text{K}_2\text{O}} - 0.25(5)C^l_{\text{P}_2\text{O}_5} - 0.15(3)\frac{C^l_{\text{Na}_2\text{O}} + C^l_{\text{K}_2\text{O}}}{P} + 0.26(3)C^l_{\text{P}_2\text{O}_5}P
\]  

(S10)

\[
C^l_{\text{MgO}} = -1.3(3) + 0.17(8)P + 6.3(4)P^{0.3} + 21.6(4)\text{Mg}^{\#}_{\text{liq}} - 0.51(1)C^l_{\text{Al}_2\text{O}_3} - 0.27(2)C^l_{\text{Na}_2\text{O}}
\]

\[
- 0.98(4)C^l_{\text{K}_2\text{O}} + 0.02(2)C^l_{\text{P}_2\text{O}_5}
\]  

(S11)

\[
C^l_{\text{FeO}} = 26.4(9) + 0.17(10)P + 8.6(5)P^{0.3} - 19.2(6)\text{Mg}^{\#}_{\text{liq}} - 0.80(2)C^l_{\text{Al}_2\text{O}_3} - 0.27(3)C^l_{\text{Na}_2\text{O}}
\]

\[
- 0.90(6)C^l_{\text{K}_2\text{O}} + 0.02(2)C^l_{\text{P}_2\text{O}_5}
\]  

(S12)

\[
T_{\text{eff,sol}} = 1196(9) + 86.4(1.5)P + 174.3\text{Mg}^{\#}_{\text{liq}} - 5.2(6)C^l_{\text{Al}_2\text{O}_3} - 7.3(7)(C^l_{\text{Na}_2\text{O}} + C^l_{\text{K}_2\text{O}})
\]

\[
- 9.6(9)C^l_{\text{P}_2\text{O}_5}
\]  

(S13)

Harzburgite melting:

\[
C^l_{\text{SiO}_2} = 52.4(1.1) - 1.2(2)P - 1.2(1.2)P^{0.3} - 3.8(7)\text{Mg}^{\#}_{\text{liq}} + 0.13(2)C^l_{\text{Al}_2\text{O}_3} - 0.65(5)C^l_{\text{Na}_2\text{O}}
\]

\[
+ 0.19(9)C^l_{\text{K}_2\text{O}} - 0.65(9)C^l_{\text{P}_2\text{O}_5} + 0.90(4)\frac{C^l_{\text{Na}_2\text{O}} + C^l_{\text{K}_2\text{O}}}{P} - 0.35(5)C^l_{\text{P}_2\text{O}_5}P
\]  

(S14)

\[
C^l_{\text{CaO}} = -0.5(9) - 1.1(2)P + 3.8(1.1)P^{0.3} + 2.6(6)\text{Mg}^{\#}_{\text{liq}} + 0.45(2)C^l_{\text{Al}_2\text{O}_3} - 0.33(4)C^l_{\text{Na}_2\text{O}}
\]

\[
- 0.20(9)C^l_{\text{K}_2\text{O}} - 0.41(9)C^l_{\text{P}_2\text{O}_5} - 0.25(4)\frac{C^l_{\text{Na}_2\text{O}} + C^l_{\text{K}_2\text{O}}}{P} + 0.48(5)C^l_{\text{P}_2\text{O}_5}P
\]  

(S15)
\[ C_{\text{MgO}}^i = -1.0(3) - 0.67(6)P + 7.6(3)P^{0.3} + 25(1)\text{Mg#}_{\text{liq}} - 0.703(8)C_{\text{Al}_2\text{O}_3}^i - 0.195(18)C_{\text{Na}_2\text{O}}^i \\
- 0.883C_{\text{K}_2\text{O}}^i - 0.03(2)C_{\text{P}_2\text{O}_5}^i \]  
(S16)

\[ C_{\text{FeO}}^i = 29.0(4) - 0.26(9)P + 8.7(5)P^{0.3} - 20.8(5)\text{Mg#}_{\text{liq}} - 0.877(12)C_{\text{Al}_2\text{O}_3}^i - 0.24(3)C_{\text{Na}_2\text{O}}^i \\
- 0.94(6)C_{\text{K}_2\text{O}}^i - 0.065(30)C_{\text{P}_2\text{O}_5}^i \]  
(S17)

\[ T_{\text{eff},\text{sol}} = 1262(8) + 66.9(9)P + 297(15)\text{Mg#}_{\text{liq}} - 13.2(4)C_{\text{Al}_2\text{O}_3}^i - 5.6(7)(C_{\text{Na}_2\text{O}}^i + C_{\text{K}_2\text{O}}^i) \\
- 13.3(9)C_{\text{P}_2\text{O}_5}^i \]  
(S18)

Note that, by default, \( C_{\text{FeO}}^i \) is instead calculated based on \( C_{\text{MgO}}^i \) and the \( K_{D_{\text{FeO}}}^{\text{liq}} \) of Toplis (2005). This option is preferred to ensure that the Mg\# of partial melts reflect the Fe–Mg equilibrium at low \( P-T \) (low FeO+MgO contents).

### S1.3 Melting progression

The equations above are used to calculate the composition of a hypothetical melt and the temperature at which this melt would be in equilibrium with mantle minerals at a given pressure \( P \). This temperature corresponds to the temperature of the effective solidus and is compared to the temperature of the mantle. If \( T_{\text{eff},\text{sol}} < T \), no melt is produced, \( P \) is decreased and those calculations are repeated. If \( T_{\text{eff},\text{sol}} > T \), melting begins but no melt is extracted until the melt fraction exceeds the critical melt fraction (\( F > \theta \)). If \( F > \theta \), a fraction of the melt (\( F - \theta \)) is extracted from the mantle, and the composition of the mantle is updated:

\[ C_{\text{stoi}} = C_{\text{stoi}}^i - C_{\text{stoi}}^i \frac{F_{\text{current}} - \theta}{1 - F_{\text{bulk}} + (F_{\text{current}} - \theta)} \]  
(S19)

with \( F_{\text{current}} \) the amount of melt present in the mantle at any given time and \( F_{\text{bulk}} \) the aggregate melt fraction (i.e., the total amount of melt produced since melting started). The first time melt is extracted, \( F_{\text{current}} = F_{\text{bulk}} \).

The composition of the aggregate melt is updated at each step:

\[ C_{\text{itol}} = C_{\text{itol}}^i \frac{(F_{\text{bulk}} - F_{\text{current}}) + C_{\text{itol}}^i (F_{\text{current}} - \theta)}{F_{\text{bulk}} - \theta} \]  
(S20)

The melt increments produced over the whole melting column are assumed to be pooled in the crust.
**S2 Summary of thermometers and barometers**

MAGMARS uses two main thermometers: equation S13 for melts in equilibrium with a lherzolite and equation S18 for melts in equilibrium with a harzburgite. Those thermometers are functions of pressure. Here, we show how accurately they can predict the experimental database. We also compare them with other liquid thermometers, including some that were previously used to constrain the temperature of martian basalts (Putirka, 2005; Lee et al., 2009; Filiberto and Dasgupta, 2011, 2015; Filiberto, 2017).

Thermometer of Putirka (2005) (A-B):

\[
\ln K_d(\text{Mg}) = -2.02 + \frac{4490.5}{T} \tag{S21}
\]

\[
\ln K_d(\text{Fe}) = -2.66 + \frac{3793.3}{T} \tag{S22}
\]

Note that \(K_d(\text{Mg})\) and \(K_d(\text{Fe})\) are the cation fraction ratios, \(\frac{X_{\text{ol}}^\text{Mg}}{X_{\text{liq}}^\text{Mg}}\) and \(\frac{X_{\text{ol}}^\text{Fe}}{X_{\text{liq}}^\text{Fe}}\), respectively, and should not be confused with the \(K_D^{\text{ol-liq}}\) introduced in Equation 1.2. Both equations have to be solved numerically to be used as a thermometer and a Matlab script is provided for this purpose.

Re-calibrated version of Putirka (2005) (A-B):

\[
\ln K_d(\text{Mg}) = -2.61 + \frac{5188.5}{T} \tag{S23}
\]

\[
\ln K_d(\text{Fe}) = -3.57 + \frac{5070.7}{T} \tag{S24}
\]

Thermometer of Putirka (2008), equation 15:

\[
T = 815.3 + 265.5Mg\#_{\text{liq}} + 15.37C_{\text{MgO}}^\text{d} + 8.61C_{\text{FeO}}^\text{d} + 6.646(C_{\text{Na}_2\text{O}} + C_{\text{K}_2\text{O}}) + 39.16P \tag{S25}
\]

Recommended \(P\)-independent thermometer (this study):

\[
T = 209.2 + 481Mg\#_{\text{liq}} + 20.86(C_{\text{FeO}} + C_{\text{MgO}}) + 3.03C_{\text{SiO}_2}^\text{d} + 10.17C_{\text{Al}_2\text{O}_3}^\text{d} + 16.97(C_{\text{Na}_2\text{O}} + C_{\text{K}_2\text{O}}) - 6.97C_{\text{P}_2\text{O}_5}^\text{d} \tag{S26}
\]

Barometer of Lee et al. (2009):

\[
P = \frac{\ln (\text{Si}_4\text{O}_8) - 4.019 + 0.0165(\text{Fe}_4\text{Si}_2\text{O}_8) + 0.0005(\text{Ca}_4\text{Si}_2\text{O}_8)^2}{-770T^{-1} + 0.0058T^{1/2}} \tag{S27}
\]

Melt composition expressed in terms of “molecular species” (8 oxygen basis):

\[
\text{Si}_4\text{O}_8 = 0.25 \times (\text{SiO}_2 - 0.5 \times (\text{FeO} + \text{MgO} + \text{CaO}) - \text{Na}_2\text{O} - \text{K}_2\text{O})
\]
\[ \text{Ti}_4\text{O}_8 = 0.25 \times \text{TiO}_2 \]
\[ \text{Al}_{16/3}\text{O}_8 = 0.375 \times (\text{Al}_2\text{O}_3 - \text{Na}_2\text{O}) \]
\[ \text{Cr}_{16/3}\text{O}_8 = 0.375 \times \text{Cr}_2\text{O}_3 \]
\[ \text{Fe}_4\text{Si}_2\text{O}_8 = 0.25 \times \text{FeO} \]
\[ \text{Mg}_4\text{Si}_2\text{O}_8 = 0.25 \times \text{MgO} \]
\[ \text{Ca}_4\text{Si}_2\text{O}_8 = 0.25 \times \text{CaO} \]
\[ \text{Na}_2\text{Al}_2\text{Si}_2\text{O}_8 = \text{Na}_2\text{O} \]
\[ \text{K}_2\text{Al}_2\text{Si}_2\text{O}_8 = \text{K}_2\text{O} \]
\[ \text{P}_{16/5}\text{O}_8 = 0.625 \times \text{P}_2\text{O}_5 \]

All oxide components are in mol% (see thermobaro.MM script)

Re-calibrated version of Lee et al. (2009) (this study):

\[
P = \frac{\ln(\text{Si}_4\text{O}_8) - 4.24 + 0.0185(\text{Fe}_4\text{Si}_2\text{O}_8) + 0.0109(\text{Ca}_4\text{Si}_2\text{O}_8) + 0.009(\text{P}_{16/5}\text{O}_8)}{-800T^{-1} + 0.0075T^{1/2}} \quad (S28)
\]

References

Agee, C. B., and D. S. Draper, Experimental constraints on the origin of Martian meteorites and the composition of the Martian mantle, *Earth and Planetary Science Letters*, 224(3-4), 415–429, doi:10.1016/j.epsl.2004.05.022, 2004.

Collinet, M., E. Médard, B. Charlier, J. Vander Auwera, and T. L. Grove, Melting of the primitive martian mantle at 0.5–2.2 GPa and the origin of basalts and alkaline rocks on Mars, *Earth and Planetary Science Letters*, 427(0), 83–94, doi:http://dx.doi.org/10.1016/j.epsl.2015.06.056, 2015.

Ding, S., R. Dasgupta, and K. Tsuno, The Solidus and Melt Productivity of Nominally Anhydrous Martian Mantle Constrained by New High Pressure-Temperature Experiments - Implications for Crustal Production and Mantle Source Evolution, *Journal of Geophysical Research: Planets*, p. e2019JE006078, doi:10.1029/2019JE006078, 2020.

Dreibus, G., and H. Wänke, Mars, a volatile-rich planet, *Meteoritics*, 20(2), 367–381, 1985.

Filiberto, J., Geochemistry of Martian basalts with constraints on magma genesis, *Chemical Geology*, 466, 1–14, doi:https://doi.org/10.1016/j.chemgeo.2017.06.009, 2017.

Filiberto, J., and R. Dasgupta, Fe2+-Mg partitioning between olivine and basaltic melts: Applications to genesis of olivine-phryic shergottites and conditions of melting in the Martian interior, *Earth and Planetary Science Letters*, 304(3-4), 527–537, doi:10.1016/j.epsl.2011.02.029, 2011.
**Figure S1:** (a) Main thermometer used by MAGMARS (harzburgite melting). (b) Thermometer of Putirka (2008), equation 15. (c) Comparison between the main MAGMARS thermometer (harzburgite melting) and equation 15 of Putirka (2008). The dashed lines represent ±50°C in (a) and (b) and ±40°C in (c).

**Figure S2:** (a) Thermometer of Putirka (2005), equations A-B. (b) Thermometer of Putirka (2005) but recalibrated using the experimental database compiled for this study. The dashed lines represent ±50°C.
Figure S3: Pressure independent thermometer (this study). The dashed lines represent ±50°C.

Figure S4: (a) Barometer of Lee et al. (2009). (b) Barometer of Lee et al. (2009) recalibrated using the experimental database compiled for this study. The dashed lines represent ±0.5 GPa.
Figure S5: (a) Barometer of Lessel and Putirka (2015), their Equation 8. (b) Thermometer of Lessel and Putirka (2015), their Equation 17. The dashed lines represent ±0.5 GPa and ±50°C.

Figure S6: Isobaric MAGMARS calculations for the Dreibus and Wänke (1985) mantle composition and comparison with experiments.
**Figure S7:** Isobaric MAGMARS calculations for the PHN1611 terrestrial peridotite *Kushiro* (1996) mantle composition and comparison with experiments
Figure S8: Melting temperature (batch) from MAGMARS and experiments for different terrestrial and martian mantle compositions. For the Earth and the KLB-1 and PHN 1611 peridotites, the experiments are from Hirose and Kushiro (1993) and Kushiro (1996), respectively. DW85 primitive mantle (Dreibus and Wänke, 1985), with the experiments of Collinet et al. (2015) at 0.5–2.0 GPa, Matsukage et al. (2013); Agee and Draper (2004) at 4.5–4.7 GPa, and this study (2.5 GPa). LF97 is the alkali-rich primitive mantle of (Lodders and Fegley, 1997) from which experiments were also performed (Ding et al., 2020). Note the decrease in solidus and melting temperature from KLB1 to LF97. PHN 1611 is more fertile (higher in K$_2$O) than KLB1. Both terrestrial compositions have higher melting temperatures due to their higher Mg#. LF97 has the lowest melting temperatures due to its high concentration in alkali elements.
Figure S9: Isobaric MAGMARS calculations highlighting the effect of phosphorus for a DW85-like composition but with no phosphorus (solid lines) and 0.32 wt.% P₂O₅ (dashed lines), double the DW85 concentration.
Figure S10: Comparison between batch and near-fractional melting calculated with Perple_X. Melting temperatures are higher for near-fractional melting due to the extraction of incompatible elements with the melt, especially at low pressure. The composition of melts are similar in both cases. Aggregate melts have slightly higher concentrations of incompatible elements (e.g. $\text{Al}_2\text{O}_3$ and $\text{Na}_2\text{O}$).
Figure S11: Comparisons MAGMARS vs. thermodynamic models
Figure S12: Comparisons MAGMARS vs. thermodynamic models
Figure S13: Comparisons MAGMARS vs. thermodynamic models
Figure S14: Comparison of the density of partial melts produced at different pressures with pMELTS and MAGMARS. For consistency and to isolate the effect of melt composition, the densities are calculated at 1300°C and 1.0 GPa regardless of the $P-T$ conditions at which the melts were initially produced, using the model of Lange and Carmichael (1990).
Filiberto, J., and R. Dasgupta, Constraints on the depth and thermal vigor of melting in the Martian mantle, *Journal of Geophysical Research: Planets*, p. 2014JE004745, doi: 10.1002/2014JE004745, 2015.

Hirose, K., and I. Kushiro, Partial melting of dry peridotites at high pressures: Determination of compositions of melts segregated from peridotite using aggregates of diamond, *Earth and Planetary Science Letters*, 114(4), 477–489, doi:http://dx.doi.org/10.1016/0012-821X(93)90077-M, 1993.

Kushiro, I., Partial Melting of a Fertile Mantle Peridotite at High Pressures: An Experimental Study Using Aggregates of Diamond, in *Earth Processes: Reading the Isotopic Code*, pp. 109–122, American Geophysical Union, doi:10.1029/GM095p0109, 1996.

Lange, R. L., and I. S. E. Carmichael, Thermodynamic properties of silicate liquids with emphasis on density, thermal expansion and compressibility, *Reviews in Mineralogy and Geochemistry*, 24(1), 25–64, 1990.

Lee, C.-T. A., P. Luffi, T. Plank, H. Dalton, and W. P. Leeman, Constraints on the depths and temperatures of basaltic magma generation on Earth and other terrestrial planets using new thermobarometers for mafic magmas, *Earth and Planetary Science Letters*, 279(1–2), 20–33, doi:http://dx.doi.org/10.1016/j.epsl.2008.12.020, 2009.

Lessel, J., and K. Putirka, New thermobarometers for martian igneous rocks, and some implications for secular cooling on Mars, *American Mineralogist*, 100(10), 2163–2171, doi:10.2138/am-2015-4732, 2015.

Lodders, K., and B. Fegley, An oxygen isotope model for the composition of Mars, *Icarus*, 126(2), 373–394, doi:10.1006/icar.1996.5653, 1997.

Matsukage, K. N., Y. Nagayo, M. L. Whitaker, E. Takahashi, and T. Kawasaki, Melting of the Martian mantle from 1.0 to 4.5 GPa, *Journal of Mineralogical and Petrological Sciences*, 108(4), 201–214, doi:10.2465/jmps.120820, 2013.

Putirka, K. D., Mantle potential temperatures at Hawaii, Iceland, and the mid-ocean ridge system, as inferred from olivine phenocrysts: Evidence for thermally driven mantle plumes, *Geochemistry, Geophysics, Geosystems*, 6(5), n/a–n/a, doi:10.1029/2005GC000915, 2005.

Putirka, K. D., Thermometers and Barometers for Volcanic Systems, *Reviews in Mineralogy and Geochemistry*, 69(1), 61–120, doi:10.2138/rmg.2008.69.3, 2008.

Toplis, M. J., The thermodynamics of iron and magnesium partitioning between olivine and liquid: criteria for assessing and predicting equilibrium in natural and experimental systems, *Contributions to Mineralogy and Petrology*, 149(1), 22–39, doi:10.1007/s00410-004-0629-4, 2005.