

Mineral-melt partitioning of redox-sensitive elements

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Abstract

Elements with variable valence state (i.e. redox-sensitive) often show contrasting mineral/melt partition coefficients as a function of oxygen fugacity (fO_2) in magmatic systems. This is because trace-element incorporation into crystal lattices depends on the charge, size, and crystal-field stabilization energy of atoms, all of which differ greatly between oxidized and reduced species of the same element. This has two critical implications: (1) petrologic/ geochemical modelling of partitioning behavior of redox-sensitive trace-elements in magmatic systems requires some knowledge of their oxidation state, and (2) the oxidation state of magmatic systems may be inferred from partitioning relations of redox-sensitive trace elements preserved in mineral and melt phases of rapidly cooled magmas. The advantage of this oxybarometric approach is that mineral/melt partitioning relations are not sensitive to late stage degassing, charge-transfer on quenching, or surficial alteration. In this chapter we discuss the theoretical treatment of experimental mineral/melt partitioning data of redox-sensitive trace elements, and review aspects concerning the partitioning behavior of well-known redox-sensitive elements, including transition metals (Ti, V, Cr, Fe), rare earth elements (Ce, Eu), U, and siderophile elements (Mo, W, Re, and platinum group elements) under planetary magmatic fO_2 conditions.

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Running head: Partitioning of redox-sensitive elements

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Abstract

Elements with variable valence state (i.e. redox-sensitive) often show contrasting mineral/melt partition coefficients as a function of oxygen fugacity (fO_2) in magmatic systems. This is because trace-element incorporation into crystal lattices depends on the charge, size, and crystal-field stabilization energy of atoms, all of which differ greatly between oxidized and reduced species of the same element. This has two critical implications: (1) petrologic/geochemical modelling of partitioning behavior of redox-sensitive trace-elements in magmatic systems requires some knowledge of their oxidation state, and (2) the oxidation state of magmatic systems may be inferred from partitioning relations of redox-sensitive trace elements preserved in mineral and melt phases of rapidly cooled magmas. The advantage of this oxybarometric approach is that mineral/melt partitioning relations are not sensitive to late stage degassing, charge-transfer on quenching, or surficial alteration. In this chapter we discuss the theoretical treatment of experimental mineral/melt partitioning data of redox-sensitive trace elements, and review aspects concerning the partitioning behavior of well-known redox-sensitive elements, including transition metals (Ti, V, Cr, Fe), rare earth elements (Ce, Eu), U, and siderophile elements (Mo, W, Re, and platinum group elements) under planetary magmatic fO_2 conditions.

1. Introduction

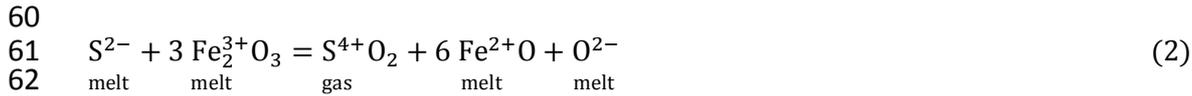
Although it is possible to determine valence-state ratios of redox-sensitive trace elements in natural and synthetic silicate glasses using techniques such X-ray absorption spectroscopy (e.g. Sutton and Newville, 2014), it may be challenging to relate these ratios to the oxygen fugacity (fO_2) of a silicate melt. This is because many elements exchange electrons with Fe, the most abundant redox-sensitive element in most magmas, upon quenching. For example, a mid-ocean ridge basaltic (MORB) melt at 1400 °C equilibrated at fO_2 equivalent to FMQ-1.7 (i.e. 1.7 log units below the fO_2 defined by the fayalite-magnetite-quartz buffer at the same temperature) has $Cr^{2+}/\Sigma Cr \sim 0.5$ (where $\Sigma Cr = Cr^{2+} + Cr^{3+}$), but upon quenching to a glass all the Cr^{2+} is converted to Cr^{3+} by reaction with Fe^{3+} (Berry et al., 2003):



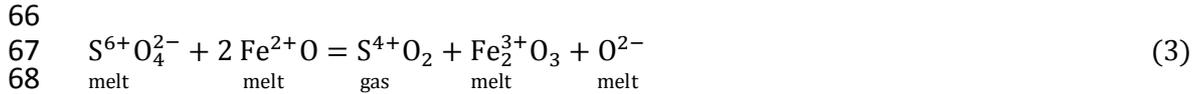
Similar reactions with Fe have been documented for Eu and Ce (Burnham and Berry, 2014; Burnham et al., 2015), but because they require only the transfer of electrons, they are probably unavoidable. Consequently, valence-state ratios of trace elements in natural Fe-bearing glasses may not be representative of their high-temperature melt chemistry.

Provided that these elements are only present at low abundances relative to Fe, the exchange minimally affects the $Fe^{3+}/\Sigma Fe$ (where $\Sigma Fe = Fe^{2+} + Fe^{3+}$) of the magma. However, this ratio can still be affected by degassing of volatile species (e.g. Burgisser and Scaillet, 2007;

55 Mathez, 1984; Sato, 1978) as well as surficial oxidation (e.g. Rhodes and Vollinger, 2005). Sulfur
 56 degassing appears to be particularly important in the Earth because of the strong capacity of S
 57 to either reduce or oxidize Fe depending on the initial oxidation state of the magma (e.g. de
 58 Moor et al., 2013; Moussallam et al., 2016). For instance, the $\text{Fe}^{3+}/\Sigma\text{Fe}$ of reduced magmas,
 59 where S^{2-} (sulfide) is the dominant melt species, may be further reduced upon degassing of SO_2 :



63
 64 whereas the $\text{Fe}^{3+}/\Sigma\text{Fe}$ of oxidized magmas, where S^{6+} (sulfate) is the dominant melt species,
 65 may be further oxidized upon degassing of SO_2 :



69
 70 While potentially significant in their own right, such effects obscure the primary
 71 magmatic redox characteristics. In contrast, high-temperature distributions of trace elements
 72 between minerals and melts (i.e. partitioning relations) are readily preserved in rocks that have
 73 cooled at moderate or rapid rates, and hence allow the petrologist to ‘see’ melt chemistry that
 74 more accurately reflects the properties of the source of a magma. Furthermore, analysis of trace
 75 elements in minerals and glasses is readily achieved using benchtop equipment, unlike some
 76 spectroscopic techniques such as XANES that require access to the handful of suitably-equipped
 77 synchrotron beamlines worldwide. Hence, mineral/melt partitioning allows more convenient
 78 insights into the redox states of magmas.

79
 80 To derive viable oxybarometers from mineral/melt partitioning, experimental data and
 81 a thermodynamic framework are required. Experimental equipment, such as gas-mixing tube
 82 furnaces, allows access to ~ 25 log units of $f\text{O}_2$ at 1400 °C; from 10 log units below the iron-
 83 wüstite buffer (IW-10) to six log units above the hematite-magnetite buffer (HM+6). For
 84 several redox-sensitive elements this is sufficient to encompass the entire transition from one
 85 valence state to another (where the redox reaction involves the exchange of 1 electron, such as
 86 Cr^{2+} to Cr^{3+} , the transition occurs over 16 log units, whereas for elements whose redox
 87 transition involves the exchange of 8 electrons, such as S^{2-} to S^{6+} , the transition occurs over 2 log
 88 units). This allows the partitioning behavior of the end-members to be determined.
 89 Nevertheless, in several cases it is not feasible to do this. For instance, where one valence state
 90 is highly volatile (e.g. Cr^{6+} or Mo^{6+}), or where an intermediate valence state always exists
 91 alongside a higher or lower valence state (e.g. V^{3+} , V^{4+} and U^{5+}), or when extreme redox
 92 conditions necessary to constrain the partitioning behavior of a valence state alone cannot be
 93 achieved in the laboratory. In such cases, it may be necessary to model the expected behavior of
 94 a valence state.

95
 96 Lattice strain theory has become one of the fundamental concepts for understanding
 97 trace element partitioning into minerals (Blundy and Wood, 1994; Wood and Blundy, 2003).
 98 The essence of the theory is that substitution of a trace element i onto a cation site in a crystal
 99 lattice is energetically neutral if the substituent ion has the same ionic radius ($r_i = r_0$) and
 100 charge as the ion it replaces, resulting in a partition coefficient D_0 . Where its ionic charge and/or
 101 radius differs, electrostatic work is done in repelling or attracting the adjacent anions (usually
 102 oxygen), and this work results in the trace element being less compatible than the one with
 103 optimal charge and radius, i.e. $D_i < D_0$. The dependence of $\log D_i$ on r_i is approximately
 104 parabolic; D_0 likewise has a quadratic dependence on atomic number, Z (Fig. 1). These two
 105 controlling variables are the reason that the various valence states of an element can have
 106 striking differences in their partitioning behavior. Consider the removal of an electron from
 107 Eu^{2+} :

108



110

111 which not only increases the charge of the cation, but also (because of its higher effective
112 nuclear charge) cause a contraction of the electronic orbitals, resulting in a decrease in ionic
113 radius. This same principle applies to other major and trace elements capable of more than one
114 valence state. As charge and ionic radius are principal controls on partitioning, it is almost
115 inevitable that different valence states of an element will behave like different elements.
116 Routine analytical techniques do not measure the speciation of redox-sensitive elements, which
117 are usually reported as total concentrations without regard for the important differences
118 between their valence states.

119

120 There are however challenges in the application of lattice strain theory to predict
121 partition coefficients. These come mainly from crystal field effects and from the multitude of
122 substitution mechanisms that are possible for a single cation in mineral sites.

123

124 Furthermore, there are important crystal-chemical implications for the fact that
125 different valence states, such as Cr^{2+} and Cr^{3+} , behave like different elements. Crystals must
126 have an overall neutral charge, hence when Cr^{2+} substitutes for Mg^{2+} in olivine, this criterion is
127 observed, with the simple substitution $\text{Mg}^{2+} \leftrightarrow \text{Cr}^{2+}$ with the endmember Cr^{2+} -olivine forming
128 according to the reaction:

129



131 melt melt olivine

132

133 which, because forsterite (Mg_2SiO_4) saturation requires $2\text{MgO} + \text{SiO}_2$ and hence a fixed
134 relationship between the activities of SiO_2 and MgO , can equivalently be written:

135

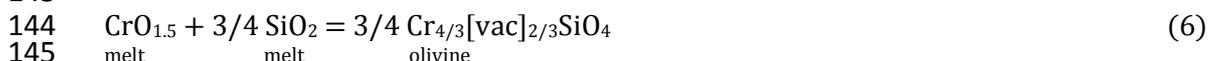


137 melt olivine. melt olivine

138

139 However, in order to substitute Cr^{3+} into the olivine structure, accommodation must be made
140 for the difference in charge. One possibility involves the creation of vacancies [vac] on the
141 octahedral site: $2 \text{Mg}^{2+} \leftrightarrow 4/3 \text{Cr}^{3+} + 2/3[\text{vac}]$. The reaction for formation of the corresponding
142 endmember component is:

143

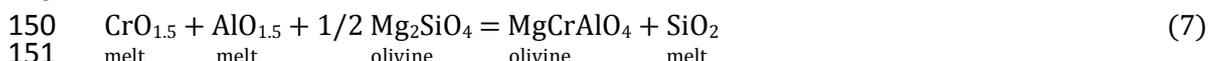


145 melt melt olivine

146

147 Another possibility involves coupled substitution with Al and Si on the tetrahedral site: $\text{Mg}^{2+} +$
148 $\text{Si}^{4+} \leftrightarrow \text{Cr}^{3+} + \text{Al}^{3+}$. The reaction for formation of the corresponding endmember component is:

149



151 melt melt olivine olivine melt

152

153 Yet a third is tied to the incorporation of water:

154



156 melt melt melt olivine

157

158 It can be seen from Equations 5-8, that these different substitutions each have their own
159 dependence on the activity of SiO_2 , and some additionally depend on the activities of Al_2O_3 and
160 H_2O . Furthermore, other components in the melt (e.g. CaO) affect the activity coefficients of CrO
161 and $\text{CrO}_{1.5}$, and accordingly the transition between Cr^{2+} and Cr^{3+} and their partition coefficients
162 are also sensitive to melt composition (Berry et al., 2006). Consequently, the examples of
163 partition coefficients quoted in this review should be taken as illustrative, rather than definitive.

209 where $K'_{10} = K_{10} / (\gamma_{M^{(x+n)^+}O_{(x+n)/2}} / \gamma_{M^{x^+}O_{x/2}})$ and the values in square brackets represent
 210 concentrations converted from mole fractions (note that the conversion factor from mole
 211 fractions to concentration cancels out between the numerator and denominator in the
 212 quotient). Taking the logarithm of this equation and re-arranging gives:
 213

$$\log \left(\frac{[M^{(x+n)^+}O_{(x+n)/2}]}{[M^{x^+}O_{x/2}]} \right) = \frac{n}{4} \log fO_2 + \log K'_{10} \quad (13)$$

214
 215

216 In the case where only two valence states are possible in the melt, by defining $\sum M = M^{x^+} +$
 217 $M^{(x+n)^+}$, one obtains:
 218

$$\frac{M^{x^+}}{\sum M} = \frac{1}{1 + 10^{(n/4 \log fO_2 + \log K'_{10})}} \quad (14a)$$

219
 220 and
 221

$$\frac{M^{(x+n)^+}}{\sum M} = \frac{1}{1 + 10^{-(n/4 \log fO_2 + \log K'_{10})}} \quad (14b)$$

222
 223

224 From this equation, it can be seen that $M^{x^+} / \sum M$ takes a sigmoidal form, constrained to vary
 225 from 0 to 1, with $\log fO_2$. The slope is defined by $n/4$ (i.e. the number of electrons in the redox
 226 reaction), and the position in $\log fO_2$ space over which the transition from one valence state to
 227 the other occurs is determined by $\log K'_{10}$ (i.e. the simplified equilibrium constant for the
 228 homogeneous redox reaction). For elements such as U and V, which have three and four
 229 possible valence states in the melt, $\sum M \neq M^{x^+} + M^{(x+n)^+}$; instead, more complicated
 230 formulations are required. In the case of Cr, which can occur as Cr^{2+} , Cr^{3+} and Cr^{6+} , there is
 231 essentially no coexistence of Cr^{2+} and Cr^{6+} , and hence Equations 14a,b are valid approximations.
 232

233
 234

235 The above treatment groups all M^{x^+} together. However, from a melt structural
 236 viewpoint, it has been argued that Fe^{3+} and Eu^{3+} occur both as dissociated cations and in
 237 complexes such as FeO_2^- and EuO_2^- (Fraser, 1975; Ottonello et al., 2001). Although it would be
 238 possible to write separate reactions for oxidation of Fe^{2+} to Fe^{3+} and FeO_2^- , the resulting
 239 equations would have an identical mathematical form to the simple model we present here. Our
 240 activity coefficients, and the use of a modified equilibrium constant $\log K'_{10}$, accommodate the
 241 terms that would be necessary to account for the speciation of the participating ions. The use of
 242 a more refined structural model may, however, prove advantageous in describing the
 243 dependence of $\log K'_{10}$ on melt composition. This is addressed in greater detail in Moretti (this
 244 volume).
 245

246
 247

2.2. Heterogeneous equilibria

248
 249

250 In the mineral (min) phase, the concentration of M will be given by the sum of all
 251 possible valence states, each with its own partitioning reaction:
 252

$$M^{x^+}_{\text{mineral}} = [M^{x^+}O_{x/2}]_{\text{melt}} + \sum [N^{x^+}O_{x/2}]_{\text{melt}} \quad (15)$$

251
 252

253 where $N^{x+}O_{x/2}$ are the ‘stoichiometric’ control needed to form the phase component in which
 254 $M^{x+}O_{x/2}$ occurs (O’Neill and Eggins, 2002). The equilibrium constant for this heterogeneous
 255 reaction is given by:

$$K_{15} = \frac{X_{M^{x+}}^{\min} \gamma_{M^{x+}O}^{\min}}{X_{M^{x+}O_{x/2}}^{\text{melt}} \gamma_{M^{x+}O_{x/2}}^{\text{melt}} \prod a_{N^{x+}O_{x/2}}^{\text{melt}}} \quad (16)$$

257
 258 where, $\prod a_{N^{x+}O_{x/2}}$ is the product of the activities of the components $N^{x+}O_{x/2}$ that make up the
 259 ‘stoichiometric control’ in the melt. The mineral/melt partition coefficient for each valence state
 260 of M is given by:

$$D_{M^{x+}} = \frac{\gamma_{M^{x+}O}^{\min}}{K_{15} \left(\gamma_{M^{x+}O_{x/2}}^{\text{melt}} \prod a_{N^{x+}O_{x/2}}^{\text{melt}} c \right)} \quad (17)$$

263
 264 where c is a constant to convert mole fractions to concentrations by weight. From this equation,
 265 it becomes clear that the mineral/melt partitioning of M will depend on the activities of the
 266 major element oxides in the system, even if they mix ideally.

267 2.3. Mineral/melt partitioning as a function of fO_2

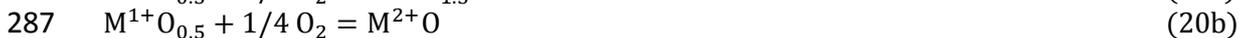
270 The mineral/melt partition coefficient of a redox-sensitive element M at constant
 271 pressure, temperature, and major element composition can be defined (alternatively to
 272 Equation 9) as the sum of the concentrations of all valence states in the mineral phase, divided
 273 by the sum of the concentration of all valence states in the melt phase:

$$D_M = \frac{\sum [M_{\min}^{x+}]}{\sum [M_{\text{melt}}^{x+}]} \quad (18)$$

274
 275 Assuming that M occurs in the valence states 1+, 2+, 3+..., this equation can be re-arranged to:

$$D_M = \frac{\frac{[M_{\min}^{1+}]}{[M_{\text{melt}}^{1+}]} + \frac{[M_{\min}^{2+}]}{[M_{\text{melt}}^{1+}]} + \frac{[M_{\min}^{3+}]}{[M_{\text{melt}}^{1+}]} \dots}{1 + \frac{[M_{\text{melt}}^{2+}]}{[M_{\text{melt}}^{1+}]} + \frac{[M_{\text{melt}}^{3+}]}{[M_{\text{melt}}^{1+}]} \dots} \quad (19)$$

281
 282 Homogeneous redox reactions can be written relating the different valence states to either the
 283 more reduced or oxidized end-members:



286 ...

287 and the ratios of the concentrations of these related to K' and fO_2 in the following way:

$$288 \quad [M_{\text{melt}}^{3+}]/[M_{\text{melt}}^{1+}] = (fO_2)^{0.5} K'_{20a} \quad (21a)$$

$$293 \quad [M_{\text{melt}}^{2+}]/[M_{\text{melt}}^{1+}] = (fO_2)^{0.25} K'_{20b} \quad (21b)$$

294 ...

295

296 Substituting Equation 21 into Equation 19 and re-arranging gives:

297

$$D_M = \frac{D^{1+} + [D^{2+}(fO_2)^{0.25}K'_{20b}] + [D^{3+}(fO_2)^{0.5}K'_{20a}] + \dots}{1 + [(fO_2)^{0.25}K'_{20b}] + [(fO_2)^{0.5}K'_{20a}] + \dots} \quad (22)$$

298

299

300 For convenience, this expression can also be written in the following form:

301

$$D_M = \frac{D^{1+} + (D^{2+} 10^{0.25 \log fO_2 + \log K'_{20b}}) + (D^{3+} 10^{0.5 \log fO_2 + \log K'_{20a}}) + \dots}{1 + (10^{0.25 \log fO_2 + \log K'_{20b}}) + (10^{0.5 \log fO_2 + \log K'_{20a}}) + \dots} \quad (23)$$

302

303

304 which then allows for the $\log fO_2$ term to be substituted more readily by oxygen fugacity terms
305 relative to buffers in the customary way (e.g. ΔFMQ , ΔIW , etc.) as a form of minimizing the effect
306 of temperature and pressure, if non-isobaric/isothermal data are considered.

307

308

309 3. Transition metals (Fe, Cr, Ti, V)

310

311 A large number of transition metals (those elements whose atoms have incompletely
312 filled $3d$ orbitals) occur in more than one valence state under magmatic fO_2 conditions (Fig. 2).
313 Amongst the redox-sensitive first-row transition metals, Fe is a major constituent of the solar
314 system while others like Ti, V, and Cr occur in minor-to-trace concentrations (Palme and O'Neill,
315 2014). Mineral/melt partition relations of transition metals are harder to constrain in
316 comparison with other trace elements as their behavior frequently does not obey Henry's Law.
317 Furthermore, their electronic configuration results in strong crystal-field effects (Burns, 1993)
318 and hence partitioning behavior is often anomalous in the context of lattice strain theory.

319

320

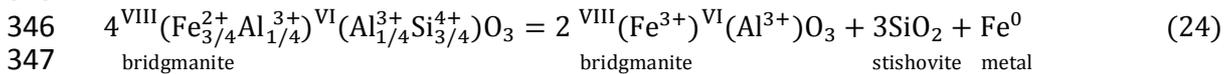
321 3.1. Iron (Fe)

322

323 As the only redox-sensitive cation that is a major element component in the solar
324 system, iron has a fundamental role in magmatic redox processes. Arguably, fO_2 in the solid
325 Earth is entirely controlled by the partitioning behavior of iron, which occurs as Fe^{3+} (ferric
326 iron), Fe^{2+} (ferrous iron) as well as Fe^0 (metallic iron). The crystal-chemical controls on Fe
327 partitioning are essential to understanding fO_2 throughout the Earth's mantle. For example, in
328 the dominant upper mantle phase olivine (Mg_2SiO_4), iron occurs almost exclusively in the
329 divalent form substituting for Mg in octahedral sites (i.e. $Fe^{3+}/\Sigma Fe \sim 0$; O'Neill et al. 1993). This
330 means that, in a system closed to oxygen, a reduced basaltic melt that crystallizes olivine will
331 become oxidized (i.e. its $Fe^{3+}/\Sigma Fe$ will increase). Conversely, an oxidized melt that crystallizes
332 magnetite ($Fe^{2+}Fe_2^{3+}O_4$, with $Fe^{3+}/\Sigma Fe \sim 0.66$) will become reduced (i.e. its $Fe^{3+}/\Sigma Fe$ will
333 decrease). Other ferromagnesian minerals in the upper mantle (e.g. pyroxenes, amphiboles,
334 garnets) will also contribute to this mass balance.

335

336 Another fundamental example that illustrates the significance of crystal-chemical
 337 controls on the partitioning behavior of Fe comes from the most abundant mineral phase in the
 338 Earth's mantle, bridgmanite (Fe,Mg)(Si,Al)O₃. Bridgmanite can contain up to 20 atomic % Fe, a
 339 substantial proportion of which is trivalent (Fe³⁺/ΣFe up to 0.6; McCammon, 1997; Frost et al.,
 340 2004). This preference of bridgmanite for ferric iron occurs because the coupled substitution of
 341 Fe³⁺ on to the VIII-fold coordinated Mg site, charge balanced by Al substitution on to the VI-fold
 342 coordinated Si site, is energetically favorable (Kesson et al., 1995). Therefore, Fe³⁺ occurs in
 343 experimentally grown Al-bearing bridgmanite even at very reducing conditions, with fO₂
 344 maintained low via a disproportionation reaction that form metallic iron (e.g. Shim et al., 2017):
 345



349 These types of disproportionation reactions are powerful in explaining the paradoxical evidence
 350 suggesting high Fe³⁺ contents in the very reduced lower mantle (McCammon, 2005).
 351

352 In other abundant ferromagnesian minerals such as pyroxenes, the incorporation of
 353 Fe²⁺ and Fe³⁺ is also strongly dictated by stoichiometric controls. The experimental data of
 354 McCanta et al. (2004), for instance, highlight significant steric constraints on the substitution of
 355 Fe³⁺ for Fe²⁺ in pigeonite (low Ca clinopyroxene) compared to augite (high Ca clinopyroxene).
 356 The limited incorporation of Fe³⁺ into pigeonite is such that even under highly oxidizing
 357 conditions only a few percent of the total Fe will be trivalent. In contrast, alkali pyroxenes like
 358 aegirine (NaFe³⁺Si₂O₆) have a strong affinity for Fe³⁺ but will not precipitate, even under
 359 oxidized conditions, unless there is sufficient Na available in the melt. As there is complete solid
 360 solution between augite and aegirine, Fe³⁺/Fe²⁺ of a pyroxene has a complex dependence on
 361 melt composition when Fe is a major element, and lies beyond the scope of this discussion. A
 362 related issue arises with the partitioning of trace elements into magnetite, which is strongly fO₂-
 363 dependent. However, this is mostly due to the stoichiometric control by the Fe³⁺/Fe²⁺ ratio of
 364 the melt rather than any variation in oxidation state of the trace elements themselves. For
 365 example, Ti becomes less compatible from FMQ to FMQ+4, well outside the range at which any
 366 Ti³⁺ is present, and this behavior is also exhibited by redox-invariant trace elements such as Zr
 367 and Hf (Siewwright et al., 2017). Hence, care should be taken when evaluating redox-related
 368 trends in Fe-rich systems to understand the factors contributing to variations in partition
 369 coefficients. The relationships described in this chapter are only applicable to trace components
 370 of melts and minerals.
 371

372 Based on experiments with simplified synthetic compositions doped with small
 373 concentrations of Fe as a function of fO₂, Mallmann and O'Neill (2009) showed that $D_{\text{Fe}^{2+}} \gg$
 374 $D_{\text{Fe}^{3+}}$ in olivine and orthopyroxene, but $D_{\text{Fe}^{2+}} < D_{\text{Fe}^{3+}}$ in clinopyroxene and spinel (Fig. 3a).
 375 Amphibole/melt data from King et al. (2000) indicates that the partitioning of both Fe³⁺ and
 376 Fe²⁺ are both close to unity, with $D_{\text{Fe}^{3+}}$ about a factor of two higher than $D_{\text{Fe}^{2+}}$ (though the range
 377 of fO₂ in this study, from FMQ-6 to FMQ+3, is not sufficient to constrain either Fe²⁺ or Fe³⁺, so
 378 the differences in D s between the two species are likely larger). The data of Martel et al. (1999)
 379 and Krawczynski et al. (2012), which examined fO₂ over an even more limited range alongside
 380 the effects of H₂O, T and P , show higher $D_{\text{Fe}}/D_{\text{Mg}}$ at more oxidizing conditions, supporting a
 381 higher compatibility for Fe³⁺ in igneous amphiboles. It should be noted in passing that
 382 amphibole Fe³⁺/ΣFe is readily altered by a redox reaction $\text{Fe}^{2+} + \text{OH}^- = \text{Fe}^{3+} + \text{O}^{2-} + \frac{1}{2} \text{H}_2$, and
 383 hence may not preserve magmatic values (Popp et al., 1995).
 384

385 In non-ferromagnesian minerals such as feldspars, iron behaves as a typical trace
 386 element. In plagioclase feldspar for instance (Fig. 3a), Fe³⁺ ($r^{\text{IV}} = 0.49 \text{ \AA}$)¹ is significantly more

¹ Ionic radii reported in this study were taken from Shannon (1976).

387 compatible than Fe^{2+} (Phinney, 1992; Wilke and Behrens, 1999) as it directly substitutes for
 388 Al^{3+} ($r^{\text{IV}} = 0.39 \text{ \AA}$) in tetrahedral sites, as opposed to Fe^{2+} which is significantly smaller ($r^{\text{VIII}} =$
 389 0.78 \AA) than the larger VIII-fold coordinated A-site ($r_0 \sim 1.2 \text{ \AA}$) (Longhi et al., 1976). In some
 390 oxidized peralkaline lavas, alkali feldspars with up to 2.8 wt.% Fe_2O_3 have been reported (Mann
 391 et al. 2006). Lundgaard and Tegner (2004) found a strong dependence of $D_{\text{Fe}^{2+}}$ and $D_{\text{Fe}^{3+}}$ on melt
 392 composition, particularly SiO_2 , but not on plagioclase composition.
 393
 394

395 3.2. Chromium (Cr)

396
 397 Chromium can exist as Cr^{2+} , Cr^{3+} and Cr^{6+} in magmas (e.g. Berry and O'Neill, 2004;
 398 Schreiber and Haskin, 1976), but direct detection of Cr^{2+} in quenched basaltic melts is hindered
 399 by an electron exchange with Fe on quenching (Reaction 1; Berry et al., 2003). Mineral-melt
 400 partitioning relations, however, have long indicated the presence of Cr^{2+} in silicate melts over
 401 typical magmatic conditions (e.g. Barnes, 1986; Hanson and Jones, 1998; Postovetov and
 402 Roeder, 2000). The divalent and trivalent species of Cr overwhelmingly dominate over typical
 403 magmatic $f\text{O}_2$ s (Fig. 2), with hexavalent Cr only becoming stable at very oxidizing conditions
 404 (usually $> \text{FMQ}+4$; Berry and O'Neill, 2004). Because such highly oxidizing conditions are
 405 uncommon in magmatic systems, and Cr^{6+} appears to be highly incompatible in high-
 406 temperature mineral phases (Mallmann and O'Neill, 2009), the presence of small amounts of
 407 Cr^{6+} in magmas should have no significant effect on the bulk partitioning of Cr (although it
 408 should be noted that the transition from Cr^{3+} to Cr^{6+} is expected to occur over a relatively
 409 narrow range of $f\text{O}_2$ given that the redox reaction involves three electrons). High-temperature
 410 experiments under the highly oxidizing conditions necessary to constrain the behavior of Cr^{6+}
 411 are limited (hexavalent Cr appears to be volatile at high temperature; Wijbrans et al., 2015), so
 412 the focus of our discussion will be on Cr^{2+} and Cr^{3+} .
 413

414 Chromium is a major constituent of many rock-forming minerals (Burns and Burns,
 415 1975), normally making end-member components where Cr^{3+} substitutes for Al^{3+} (and Fe^{3+}).
 416 Some of these substitutions are complete solid solutions, most notably between spinel
 417 ($\text{MgAl}_2^{3+}\text{O}_4$) and magnesiochromite ($\text{MgCr}_2^{3+}\text{O}_4$), and between grossular ($\text{Ca}_3\text{Al}_2^{3+}\text{Si}_3\text{O}_{12}$) and
 418 uvarovite ($\text{Ca}_3\text{Cr}_2^{3+}\text{Si}_3\text{O}_{12}$). In most cases, Cr^{3+} only occurs in octahedral coordination, as a
 419 result of its large crystal-field stabilization energy (CFSE; Burns, 1993). The ionic radius of Cr^{3+}
 420 is also significantly larger than Al^{3+} (and hence Si^{4+}), making it difficult for tetrahedral
 421 substitution. Incorporation of Cr^{2+} is unfavored due to its electronic configuration, which in
 422 octahedral coordination induces a destabilizing crystal-field energy and Jahn-teller distortions
 423 (Burns, 1975). Hence, Cr^{2+} can substitute for Mg^{2+} but only to a limited extent, and endmember
 424 Cr^{2+} silicates do not occur in nature (Li et al., 1995).
 425

426 The importance of Cr to mantle phase equilibria (e.g. Liu and O'Neill, 2004), the link
 427 between chromite and precious metals (e.g. platinum group elements, PGE) in layered
 428 intrusions (e.g. Barnes, 1986), and its utility as an oxybarometer in magmatic systems (e.g.
 429 Postovetov and Roeder, 2000) has prompted numerous studies aimed at determining Cr
 430 solubility in silicate melts (e.g. Roeder and Reynolds, 1991) and its partitioning between
 431 associated phases. Here we focus on minerals where Cr substitutes as a trace component, as this
 432 allows us to constrain the effect of $f\text{O}_2$ on the partitioning behavior of both Cr^{2+} and Cr^{3+} .
 433

434 Amongst minerals where Cr substitutes as trace element, olivine has the largest
 435 available experimental partitioning dataset as a function of $f\text{O}_2$. In olivine, Cr^{2+} ($r^{\text{VI}} = 0.80 \text{ \AA}$) is
 436 incorporated by direct replacement of Mg^{2+} ($r^{\text{VI}} = 0.72 \text{ \AA}$) in octahedral sites without significant
 437 lattice distortion (Reaction 5; Jollands et al., 2018). Substitution of Cr^{3+} ($r^{\text{VI}} = 0.615 \text{ \AA}$) into
 438 olivine octahedral sites, on the other hand, creates charge imbalance that can be accounted for
 439 either by octahedral vacancies (Reaction 6; Hanson and Jones, 1998; Jollands et al., 2018; Papike

440 et al., 2005), or by coupled substitution with Al³⁺ and Si⁴⁺ in tetrahedral sites (Reaction 7;
441 Jollands et al., 2018).

442

443 Olivine/melt partitioning studies from Schreiber and Haskin (1976) and Hanson and
444 Jones (1998) in the CaO-MgO-Al₂O₃-SiO₂ (CMAS) show a systematic dependence of $D_{Cr^{3+}}$ on melt
445 composition (Fig. 4), though the dependence is not straightforward, suggesting that both
446 substitution mechanisms described above likely operate. Hanson and Jones (1998), however,
447 were able to parameterize $D_{Cr^{3+}}$ between olivine and melt using the universal melt descriptor
448 NBO/T ($D_{Cr^{3+}} = -0.39 \text{ NBO}/T + 1.29$). For Cr²⁺, however, partitioning between olivine and melt
449 appears to be independent of melt composition, but varies slightly with temperature (Hanson
450 and Jones, 1998). More often than not, $D_{Cr^{2+}}$ is nearly equal to $D_{Cr^{3+}}$ between olivine and melt
451 (including for basaltic compositions), so that effectively no change in bulk Cr partitioning is
452 detected as a function of *f*O₂ (Mallmann and O'Neill, 2009; Hanson and Jones, 1998; Fig. 3b).
453 Direct Cr valence measurements in olivine by XANES have confirmed these findings (Bell et al.
454 2014).

455

456 Mallmann and O'Neill (2009) found that Cr²⁺ is similarly partitioned amongst olivine,
457 ortho- and clinopyroxene, but Cr³⁺ is significantly more compatible in pyroxenes, particularly
458 clinopyroxene (Fig. 3b). This is not surprising given that Cr³⁺ ($r^{VI} = 0.615 \text{ \AA}$) has an ionic radius
459 that is similar to the ideal strain-free octahedral M1 site of clinopyroxenes for trivalent cations
460 ($r_0^{3+,M1} \sim 0.668 \text{ \AA}$; Hill et al., 2000). Cr²⁺ ($r^{VI} = 0.80 \text{ \AA}$), on the other hand, is significantly larger
461 than the ideal M1 site of clinopyroxenes for divalent cations ($r_0^{2+,M1} \sim 0.69 \text{ \AA}$; Hill et al., 2000).
462 Mallmann and O'Neill (2009) noted a significant difference between the partitioning of Cr³⁺
463 between clinopyroxene and melt obtained for their compositions V1 and V7 (which are
464 identical except for the Cr-doping levels), suggesting that Cr³⁺ does not behave according to
465 Henry's Law in the concentration range of these experiments (> 1.5 wt.% Cr₂O₃), probably
466 because of the strong coupling of Cr³⁺ with Al³⁺ in pyroxenes (Klemme and O'Neill, 2000).
467 Chromium is noticeably more compatible in augite compared to pigeonite and enstatite, given
468 the availability of cations (Al³⁺ and Na⁺) for charge balancing the substitution of Cr³⁺ for Mg in
469 octahedral sites (Karner et al., 2007; Mallmann and O'Neill, 2009). Interestingly, Cr²⁺ appears to
470 be dominant species entering the structure of lower mantle phases such as ferropericlase and
471 bridgmanite (Eeckhout et al., 2007). Neither Cr³⁺ nor Cr²⁺ appears to be easily incorporated
472 into plagioclase (Aigner-Torres et al., 2007) as Cr³⁺ is too large for the tetrahedral site, and Cr²⁺
473 too small for the large distorted VIII-fold coordinated site.

474

475 Finally, we note that Cr is a well-known outlier in lattice-strain parabolas due to large
476 crystal-field effects, so the prediction of Cr²⁺ and Cr³⁺ partitioning using lattice strain models
477 should be done with great caution.

478

479

480 3.3. Titanium (Ti)

481

482 Titanium primarily occurs as Ti⁴⁺ over most of the range of *f*O₂s relevant to planetary
483 scientists, but spectroscopic studies indicate that under very reducing conditions (i.e. below
484 FMQ-4) a substantial fraction of Ti occurs as Ti³⁺ (Schreiber, 1977). The transition from Ti³⁺ to
485 Ti⁴⁺ appears to be strongly dependent on melt composition (Borisov, 2012; Fig. 2). Hence,
486 although the partitioning of Ti is a poor oxybarometer for terrestrial magmatism, it is markedly
487 more relevant to understanding the geochemical evolution of the early solar nebula and
488 reduced bodies such as the Moon (Fig. 2).

489

490 Hibonite (CaAl₁₂O₁₉) occurs in the calcium aluminum-rich inclusions (CAIs) of
491 chondritic meteorites. It is one of the earliest phases to condense from the solar nebula, and can
492 contain up to 8 wt% TiO₂ (Allen et al., 1978). The blue color of many hibonite grains has been
493 shown to be due to the presence of Ti³⁺ (Burns and Burns, 1984; Sutton et al., 2017),

494 incorporated as a $\text{CaTi}_2^{3+}\text{Al}_{10}\text{O}_{19}$ component, but Ti^{4+} can also be accommodated in hibonite via
 495 a coupled substitution with Mg, as a $\text{CaMg}_2\text{Ti}_2^{4+}\text{Al}_8\text{O}_{19}$ component (Doyle et al., 2014). The
 496 coupled nature of the substitution of Ti^{4+} means that the availability of Mg exerts a strong
 497 control on $\text{Ti}^{3+}/\Sigma\text{Ti}$ (where $\Sigma\text{Ti} = \text{Ti}^{3+} + \text{Ti}^{4+}$) in hibonite; since it is often unclear which phases
 498 are in equilibrium in CAIs, this oxybarometer may have limited utility (Berry et al., 2017).
 499

500 Ti^{3+} is more compatible than Ti^{4+} in the majority of phases, including olivine, diopside,
 501 orthopyroxene and anorthite (Mallmann and O'Neill, 2009; Peters et al., 1995; Fig. 3c). Ti K-
 502 edge XANES measurements of both synthetic and natural lunar basalt samples have revealed
 503 that Ti^{3+} is enriched in ilmenite, armalcolite and clinopyroxene at $f\text{O}_2$ of about FMQ-4.5 (Leitzke
 504 et al., 2018; Simon and Sutton, 2018). However, as it is most compatible in clinopyroxene,
 505 essentially all research has focused on this phase. Indeed, a Ti^{3+} -pyroxene (grossmanite,
 506 $\text{CaTi}^{3+}\text{AlSiO}_6$) has been reported from the Allende meteorite, and $\text{NaTi}^{3+}\text{Si}_2\text{O}_6$ has been
 507 synthesized experimentally (Ohashi et al., 1982; Ma and Rossman, 2009). Although quantitative
 508 oxybarometry is hampered by multiple sources of uncertainty, including the equilibrium
 509 mineral assemblage (as with hibonite), the mixing relations of Ti-bearing pyroxenes and the
 510 temperature of formation, decreasing $\text{Ti}^{3+}/\Sigma\text{Ti}$ from core to rim of some CAIs suggests
 511 increasingly oxidizing conditions in the solar nebula over a period of up to 300 kyr, perhaps by
 512 as much as 6 – 7 log $f\text{O}_2$ units (Young et al., 2012). However, in some CAIs an oxidation trend,
 513 followed by a more reduced rim, has been observed. This is attributed to incompatible Ti^{4+}
 514 concentrating in the melt phase, which was isolated from the solar nebula by a thick melilite
 515 mantle, followed by late-stage ingress of nebular gases to reduce the residual melt prior to
 516 crystallization of the rims (Papike et al., 2016). Ti^{3+} has also been observed in lunar pyroxenes
 517 (Burns et al., 1972), but is thought to be a minor component even on such a reduced planetary
 518 body (Papike et al., 2005). Terrestrial magmas generally form at far more oxidizing conditions
 519 and Ti is expected to occur exclusively as Ti^{4+} (Fig. 2). Pyroxenes in the native iron-bearing
 520 lavas of Disko Island, Greenland, have been suggested to contain some Ti^{3+} (Pedersen, 1981),
 521 but this is speculative and may not be compatible with the rather high Fe contents.
 522

523

524 3.4. Vanadium (V)

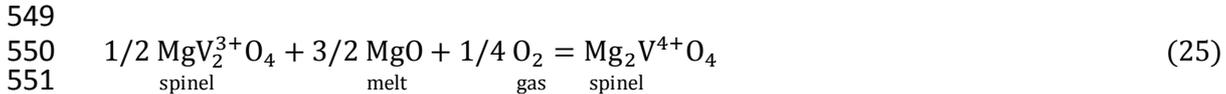
525

526 Vanadium is the redox-sensitive trace element that occurs in the largest number of
 527 valence states at typical magmatic $f\text{O}_2$ s (Fig. 2), namely V^{2+} , V^{3+} , V^{4+} and V^{5+} (Borisov et al.,
 528 1987; Schreiber et al., 1987; Sutton et al., 2005). Given the plethora of valence states with
 529 theoretically contrasting geochemical behaviors, significant effort has been made to
 530 characterize the effect of $f\text{O}_2$ and other variables such as pressure, temperature, mineral and
 531 melt composition on the partitioning of V in silicate/oxide phases relevant to mantle melting
 532 and magmatic differentiation (e.g. Arató and Audétat, 2017; Canil 1997; 1999; 2002; Canil and
 533 Fedortchouk, 2000; 2001; Karner et al., 2008; Laubier et al., 2014; Mallmann and O'Neill, 2009;
 534 2013; Papike et al., 2015; Righter et al. 2006a,b; 2011; Shearer et al., 2006; Sievwright et al.,
 535 2017; Toplis and Corgne 2002; Wijbrans et al., 2015). The end goal of many of these studies is to
 536 calibrate the sensitivity of V partitioning to $f\text{O}_2$ as an oxybarometer.
 537

538

539 Under terrestrial magmatic $f\text{O}_2$ conditions (Fig. 2), V^{3+} and V^{4+} are the dominant valence
 540 states, and spinel the major mineral group host. In pure magnetite ($\text{Fe}^{2+}\text{Fe}_2^{3+}\text{O}_4$), where the
 541 stoichiometry fixes $\text{Fe}^{3+}/\text{Fe}^{2+}$ at 2, incorporation of vanadium occurs via an $f\text{O}_2$ -independent
 542 electron-exchange reaction: $\text{Fe}^{3+} + \text{V}^{3+} = \text{Fe}^{2+} + \text{V}^{4+}$ (O'Neill and Navrotsky, 1984). However,
 543 experimental studies of spinel group minerals have invariably shown strong dependence of
 544 vanadium partitioning with $f\text{O}_2$ both in Fe-free and Fe-bearing systems (e.g. Arató and Audétat,
 545 2017; Canil et al., 1999; Mallmann and O'Neill, 2009; Papike et al., 2015; Righter et al., 2006b;
 546 Sievwright et al., 2017; Toplis and Corgne, 2002). Importantly, because Fe is an essential
 component of magnetite, the partitioning of non-redox variable elements depends on $f\text{O}_2$

547 (Siewwright et al., 2017); this is not the case for Fe-poor spinels. For example, the ratio V^{3+}/V^{4+}
 548 of pure spinel ($MgAl_2O_4$) is controlled by the reaction:



551
 552
 553 It is therefore unsurprising that the bulk partitioning of V between spinel group minerals and
 554 silicate melt is a strong function of fO_2 as well as composition (mineral and melt). Experimental
 555 data shows that $D_{V^{3+}} \gg D_{V^{4+}} > D_{V^{5+}}$ between spinel and silicate melt (Fig. 3d), and that the bulk
 556 partitioning of vanadium for Cr-spinels is higher than that for Al-spinels (Canil, 2002; Mallmann
 557 and O'Neill, 2009). The negative correlation between Al and V^{3+} in experimentally grown
 558 spinels provides strong evidence that the preferred mode of vanadium substitution into spinel
 559 is by direct replacement of Al in octahedral sites (Balan et al., 2006; Canil, 2002; Righter et al.,
 560 2006b). However, the details of the various possible substitution mechanisms are complex,
 561 given the structure of spinels and the multivalent nature of their cations (e.g. Papike et al.,
 562 2015). Melt composition also appears to play a significant role, as demonstrated by V
 563 partitioning experiments between magnetite and silicic melts (Arató and Audétat, 2017).

564
 565 Owing to its high charge and small ionic radius, V^{5+} is highly incompatible in all rock-
 566 forming minerals. When sufficiently oxidized conditions are achieved, the partitioning of V^{5+}
 567 can be obtained, though this value is typically $\ll 0.01$ (e.g. Mallmann and O'Neill, 2009). V^{4+}
 568 has partition coefficients that fall between V^{3+} and V^{5+} (Fig. 3d) and therefore is much harder to
 569 constrain. The ionic radius of V^{4+} is similar to that of Ti^{4+} in both octahedral and tetrahedral
 570 coordination ($r^{VI} = 0.58$ and 0.605 Å; $r^{IV} = 0.46$ and 0.42 Å; for V^{4+} and Ti^{4+} respectively), and
 571 both have similar electronic configurations and thus no significant difference in CFSE, so an
 572 approximation can be obtained using lattice strain theory. In connection with that, we note that
 573 $D_{V^{4+}}$ for pyroxene and olivine obtained by Mallmann and O'Neill (2009) are about an order of
 574 magnitude higher than what is inferred from lattice strain parabolas of other tetravalent
 575 cations. For some minerals, constraining the partitioning of V^{3+} may also be difficult when it
 576 shows compatibility between V^{2+} and V^{4+} (e.g. olivine or orthopyroxene). For clinopyroxene
 577 though, V^{3+} is clearly more compatible than both V^{2+} and V^{4+} , and its partitioning readily
 578 determined (Fig. 3d). Partition coefficients for V^{2+} can be obtained if sufficiently reduced
 579 experimental data is available, such as those reported by Mallmann and O'Neill (2009), but
 580 extrapolation from lattice strain parabolas should be avoided since V^{2+} has the same electronic
 581 configuration as Cr^{3+} and therefore large CFSE in octahedral coordination (Burns, 1993).

582
 583 For mineral phases that only crystallize at high pressure from natural magmas,
 584 partitioning relations are harder to constrain as a function of fO_2 since this intensive parameter
 585 needs to be buffered externally using double-capsule techniques or internally using low-
 586 solubility solid metal-oxide. Righter et al. (2011) extended the fO_2 conditions obtained by
 587 Mallmann and O'Neill (2007, 2009) to constrain the $D_{V^{3+}}$ in garnet at ~ 10 . The compatibility
 588 order of vanadium valences in garnet is $D_{V^{3+}} \gg D_{V^{4+}} > D_{V^{5+}}$, with values comparable to
 589 clinopyroxene (Mallmann and O'Neill, 2009). XANES studies by Righter et al. (2011), however,
 590 suggest that small amounts of V^{2+} are incorporated into garnet, even under relatively oxidizing
 591 conditions.

592
 593 Because of the complexities in developing a thermodynamic partitioning model that
 594 involves four different oxidation states, all of which likely varying differently with
 595 compositional parameters, Mallmann and O'Neill (2013) developed an empirical calibration of
 596 V partitioning between olivine and melt over a restricted range of fO_2 s (from FMQ-4 to FMQ+4),
 597 where $\log D_V$ vary more or less linearly with $\log fO_2$, accounting for the effects of olivine and
 598 melt composition. Similarly, Arató and Audétat (2017) developed an empirical calibration for
 599 magnetite. It would be useful to develop similar models for other common magmatic minerals,
 600 but this requires more extensive experimental databases since substitution of different V

601 valence states in minerals like pyroxene and garnet are more complex than in olivine and
602 magnetite.

603

604

605 4. Rare earths (Ce and Eu)

606

607 The rare earth elements (REE), including Y, have favorable characteristics for examining
608 redox processes. Apart from Ce and Eu, the REE exist exclusively in the trivalent state and their
609 ionic radii vary smoothly from La to Lu (with Y somewhere between Ho and Er). This provides
610 excellent coverage for fitting a lattice strain parabola to the REE³⁺; anomalies relative to this
611 parabola are readily quantifiable, and often visually striking. Zircon is notable because its REE
612 patterns commonly feature both Ce and Eu anomalies (Fig. 5). They are typically quantified as
613 Ce/Ce* and Eu/Eu*, where Ce* and Eu* are the values that would be predicted by interpolation
614 or extrapolation. The simplest is the geometric mean of La and Pr (i.e., a linear interpolation on
615 a logarithmic axis):

616

$$617 \text{Ce}^* = (\text{La}_N \times \text{Pr}_N)^{0.5} \quad (26)$$

618

619 where 'N' denotes normalization to CI chondrite, whole rock or matrix glass. La is often below
620 the limit of detection for analyses of zircon, and so extrapolation from Pr-Nd-Sm may be used
621 instead. Alternatively, polynomial fits to the REE pattern may be used (e.g. O'Neill, 2016;
622 Burnham and Berry, 2017). There are systematic deviations between the methods, so values
623 taken from different sources may need recalculation before being compared.

624

625 Cerium exists as both Ce³⁺ and Ce⁴⁺ in magmas. However, at typical magmatic fO₂ the
626 proportion of Ce⁴⁺ is small (Fig. 2), with Ce⁴⁺/ΣCe ~ 0.001 (Burnham and Berry, 2012; 2014).
627 At such low levels, it can be seen from Fig. 6a that where Ce⁴⁺ is less compatible than Ce³⁺ there
628 is no perceptible effect on the bulk partitioning of Ce: even in the most extreme case, i.e. $D_{\text{Ce}^{4+}} =$
629 0 , D_{Ce} would drop by 0.1 %, an amount that is far less than the achievable analytical accuracy. By
630 a similar line of logic, it can be calculated that in order for a mineral in a terrestrial magma to
631 develop a noticeable Ce anomaly, e.g. Ce/Ce* = 1.1, $D_{\text{Ce}^{4+}}$ must be greater than $D_{\text{Ce}^{3+}}$ by a factor of
632 100. For minerals with a steep or strongly curved REE patterns, or larger analytical
633 uncertainties, even a 10 % excess of Ce might not be a convincing anomaly.

634

635 Few minerals have a strong affinity for Ce⁴⁺. The most notable, zircon, will be addressed
636 in detail below. This is a favorable substitution on account of the identical 4+ charge of Zr, and a
637 similarly large ionic radius ($r^{\text{VIII}} = 0.84$ and 0.97 Å for Zr and Ce⁴⁺ respectively). Positive Ce
638 anomalies have been reported from other Zr minerals such as baddeleyite (Schärer et al., 2011),
639 wadeite (Jaques, 2016), elpidite and armstrongite (Gerdes et al., 2017). Thomson et al. (2016)
640 reported both positive and negative Ce anomalies for inclusions in diamonds, but the precise
641 interpretation of these grains is complicated by their polymineralic nature and a lack of
642 experimental partitioning data.

643

644 Zircon is the commonest of these, and has attracted the most experimental work
645 (Burnham and Berry, 2012; Trail et al., 2011; 2012). The most recent calibration of the Ce-in-
646 zircon oxybarometer uses lattice strain theory to infer $D_{\text{Ce}^{3+}}$ and $D_{\text{Ce}^{4+}}$, which allow Ce⁴⁺/ΣCe in
647 the melt to be calculated and the corresponding fO₂ deduced (Smythe and Brenan, 2016). This
648 approach avoids the challenge of growing analyzable well-equilibrated zircon crystals, but there
649 is presently a discrepancy between the two available models for Ce speciation in silicate melts
650 (Burnham and Berry, 2014; Smythe and Brenan, 2015) and hence further work is required to
651 develop a robust version of this oxybarometer.

652

653 For other minerals, there has been no systematic study of the effect of fO₂, but this can
654 be predicted by a combination of lattice strain modelling (and for minerals with more than one

655 cation site, whose 4+ parabolas are underconstrained, by making the approximation $D_{Ce^{4+}} =$
 656 $D_{U^{4+}}$ and $Ce^{4+}/\Sigma Ce$ estimated from the optical basicity of the melt composition (Burnham and
 657 Berry, 2014). It can be seen that Ce^{4+} is less compatible than Ce^{3+} in Zr-free minerals and that,
 658 as explained above, D_{Ce} does not change appreciably in the fO_2 range of natural magmas.
 659

660 Although the foregoing discussion considered mineral/melt partitioning, it is not always
 661 possible to consider mineral compositions in terms of mineral/melt partition coefficients. In
 662 particular, plutonic rocks do not retain any melt fraction, but even volcanic rocks often contain a
 663 significant cargo of crystals that are substantially older and may not be in equilibrium with their
 664 present host (e.g. Murphy et al., 1998). In the case of Ce, it is possible to use the bulk rock or
 665 even CI chondrite as a normalization factor, because magma compositions with Ce anomalies
 666 are uncommon. It is often stated that subduction zone magmas can have $Ce/Ce^* < 1$, indicating
 667 the incorporation of Ce-depleted sediments. As highlighted by O'Neill (2016), however, many
 668 reported examples of Ce anomalies in lavas are poorly constrained. Relatively small negative Ce
 669 anomalies, $Ce/Ce^* \sim 0.9$, are known in a variety of subduction settings (e.g. Hastie et al., 2009;
 670 Woodhead, 1989). As zircon Ce anomalies are typically one or two orders of magnitude larger,
 671 failure to account for a pre-existing Ce anomaly in the magma results in an underestimate of log
 672 fO_2 of < 0.2 . The same is not true for Eu.
 673

674 Europium can exist as Eu^{2+} and Eu^{3+} in magmas. The transition occurs most rapidly
 675 below the FMQ buffer, a range over which $Fe^{3+}/\Sigma Fe$ changes more slowly, and hence it is
 676 potentially a useful oxybarometer for reduced magmas (Fig. 2). The crystal chemistry of Eu^{2+}
 677 ($r^{VI} = 1.17 \text{ \AA}$) can be thought of as similar to Sr^{2+} ($r^{VI} = 1.18 \text{ \AA}$), and indeed there is even a Eu^{2+}
 678 feldspar (Kimata, 1988). In contrast, Eu^{3+} behaves as a typical middle REE. Accordingly, $D_{Eu^{2+}} >$
 679 $D_{Eu^{3+}}$ for plagioclase and alkali feldspars and åkermanitic melilite (Kuehner et al., 1989), but for
 680 most other minerals $D_{Eu^{3+}} > D_{Eu^{2+}}$ (Fig. 6b). Plagioclase feldspar and clinopyroxene are by far the
 681 most-studied in terms of Eu partitioning as a function of fO_2 and the following discussion will be
 682 restricted to these, although phases such as titanite, with $D_{Eu^{3+}}/D_{Eu^{2+}} \approx 2500$ may be important
 683 to consider for some igneous systems (Loader et al., 2017).
 684

685 There are various models that allow plagioclase/melt $D_{Eu^{2+}}$ and $D_{Eu^{3+}}$ to be predicted
 686 using lattice strain theory (Aigner-Torres et al., 2007; Dohmen and Blundy, 2014).
 687 Corresponding models for clinopyroxene and other minerals have not been developed, and even
 688 in the case of plagioclase, Eu oxybarometry will be more precise where measured values of
 689 partition coefficients for proxy elements are used for $D_{Eu^{2+}}$ and $D_{Eu^{3+}}$. Rearranging Equation 9, we
 690 can approximate $Eu^{3+}/\Sigma Eu$ by:
 691

$$692 \quad Eu^{3+}/\Sigma Eu = (D_{Eu} - D_{Sr}) / (D_{MREE} - D_{Sr}) \quad (27)$$

693
 694 where D_{Eu} is the observed Eu partition coefficient and D_{MREE} is an estimate of $D_{Eu^{3+}}$, i.e. D_{Sm} , D_{Gd} ,
 695 or preferably $(D_{Sm} \times D_{Gd})^{0.5}$. For values of $Eu^{3+}/\Sigma Eu$ in the range 0.1 to 0.9, the fO_2 can be
 696 calculated by rearranging the general equation from Burnham et al. (2015):
 697

$$698 \quad \log fO_2 = 40.4 - 25640/T - 56.8\Lambda - 4 \log (\Sigma Eu/Eu^{3+} - 1) \quad (28)$$

699
 700 where T is temperature in Kelvin, and Λ is optical basicity, which can be calculated as:
 701

$$702 \quad \Lambda = \frac{\frac{SiO_2}{62.6} + \frac{TiO_2}{36.3} + \frac{Al_2O_3}{56.6} + \frac{FeO}{55.8} + \frac{MgO}{51.7} + \frac{CaO}{56.1} + \frac{Na_2O}{53.2} + \frac{K_2O}{67.3}}{\frac{SiO_2}{30.04} + \frac{TiO_2}{39.9} + \frac{Al_2O_3}{34} + \frac{FeO}{55.8} + \frac{MgO}{40.3} + \frac{CaO}{56.1} + \frac{Na_2O}{62} + \frac{K_2O}{94.2}} \quad (29)$$

704 with all oxide components in wt. %, and no distinction drawn between Fe²⁺ and Fe³⁺. The effect
 705 of H₂O was not included in the optical basicity model of Duffy (1993) and hence is not
 706 considered here. At natural concentrations, MnO and P₂O₅ have negligible influence (e.g. 0.5
 707 wt.% P₂O₅ would change log *K'* for Eu²⁺-Eu³⁺ by ~0.02).

708

709 With values of Eu³⁺/ΣEu outside the range 0.1 to 0.9, the reliability of this type of
 710 oxybarometer is compromised by the shallow slope of M^{x+}/ΣM against fO₂. An error of ±0.03
 711 propagates to more than a log unit of fO₂. A key challenge in using Eu partitioning to determine
 712 fO₂ is the strong dependence of log *K'* on melt composition (Aigner-Torres et al., 2007; Burnham
 713 and Berry, 2015). The relationship given in Equation 28 was derived for alkali-free
 714 compositions, though experimental data for Na- and K-bearing compositions in the literature
 715 are generally consistent with this relationship. Moreover, although it lies outside the scope of
 716 this review, we note that in hydrothermal fluids the Eu²⁺-Eu³⁺ equilibrium is 12 times more
 717 sensitive to pH than to log fO₂, and hence this variable must also be considered in low-
 718 temperature systems (Brugger et al., 2008).

719

720

721 5. Uranium (U)

722

723 Although it has been known for many years that U can exist as U⁴⁺, U⁵⁺ and U⁶⁺ in
 724 silicate melts (e.g. Calas 1979; Schreiber 1983), this fact has often been rather under-
 725 appreciated in the geochemical literature. The radioactivity of U generates heat as well as a
 726 series of decay products culminating in Pb, and hence understanding the geochemistry of U is
 727 particularly important. Interpretation of U-series disequilibrium depends on accurate partition
 728 coefficients for U and Th (as well as other elements in the series). Because of potential
 729 differences in redox between mid-oceanic ridges and subduction zone settings, for example, it is
 730 essential to consider the role of fO₂ in controlling the partitioning of U relative to Th.

731

732 Partitioning of U as a function of fO₂ has only been studied for a handful of minerals (Fig.
 733 7). Despite the importance of garnet in influencing U-series disequilibrium (Beattie, 1993), little
 734 is known about the how its partitioning of U might be affected by fO₂. For all minerals studied to
 735 date, it appears that *D*_{U⁴⁺} > *D*_{U⁵⁺} > *D*_{U⁶⁺}. As with the intermediate oxidation states of V, the
 736 partitioning of U⁵⁺ is difficult to constrain without independent spectroscopic measurements of
 737 U⁵⁺/ΣU in the same melt composition. An additional complication arises because U⁶⁺ is volatile
 738 and is gradually lost from 1 atm experiments, leading to uncertainty on *D*_{U⁶⁺} because diffusion is
 739 unlikely to occur fast enough to keep crystals in equilibrium with the changing melt
 740 composition (Burnham and Berry, 2012).

741

742 Although there are insufficient data to develop a general expression for Uⁿ⁺/ΣU as a
 743 function of melt composition, temperature, pressure and fO₂, it is known that higher oxidation
 744 states are favored by lower temperatures and melts with higher optical basicities (Schreiber,
 745 1983; Halse, 2014). Notably, U⁵⁺ appears to have a reduced stability field with increasing
 746 pressure (Halse, 2014): at higher pressure it appears that U⁴⁺ oxidizes directly to U⁶⁺, leading to
 747 a sharper change in partition coefficients as a function of fO₂ (Mallmann et al., 2016).

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749 U-Pb dating of zircon is widely regarded as one of the most reliable techniques for
 750 determining rock ages. For zircons younger than a few million years old, it becomes important
 751 to consider the contribution of ²³⁰Th to the ingrowth of ²⁰⁶Pb, because minerals for which U is
 752 less compatible than Th will develop a ²⁰⁶Pb excess, and vice versa. Correcting for these
 753 consequences of U-series disequilibrium necessitates estimates of *D*_{Th}/*D*_U. U is more compatible
 754 than Th under reducing conditions, but much less compatible at high fO₂: the variation in
 755 terrestrial magmatic redox conditions equates to a ~ 100 kyr range of possible age corrections
 756 (McLean et al., 2011).

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6. Siderophile elements (Mo, W, Re, Pt group elements)

Siderophile elements (i.e. those with strong tendency to alloy with Fe) are important tracers of planetary differentiation owing to their propensity of fractionating into planetary cores. Because of this, a great deal of effort has been made to investigate the behavior of these elements during core formation and to determine their valence state in silicate melts.

6.1. Molybdenum (Mo), tungsten (W) and rhenium (Re)

Mo, W and Re are known to occur in 4+ and 6+ valence states at fO_2 s relevant to planetary magmatism (Cottrell et al., 2009; Ertel et al., 2001; Holzheid et al., 1994; O'Neill & Eggins, 2002; O'Neill et al., 2008; Righter et al., 2016; Wade et al., 2012; 2013). While the hexavalent state of Mo, W and Re is dominant over most planetary conditions (Fig. 2), the mineral/melt partitioning behavior of these elements is strongly controlled by their tetravalent state (Fig. 8).

A common feature of the partitioning behavior of Mo, W and Re is that their 4+ valence state is significantly more compatible in silicates and oxides (e.g. spinel) than their 6+ valence state, the latter of which being almost perfectly incompatible in crystalline phases. The relative compatibility in silicates and oxides tends to follow the order $D_{\text{cpx}} > D_{\text{opx}} > D_{\text{olv}} > D_{\text{spl}} \sim D_{\text{plg}}$, independently of valence state (Fonseca et al., 2014; Leitzke et al., 2016; 2017; Mallmann and O'Neill, 2007; Wijbrans et al., 2015). For all three elements, the difference between the mineral/melt partition coefficient of their reduced and oxidized species spans three to five orders of magnitude (Fig. 8). The difference in partition coefficients between the valence states is most pronounced for clinopyroxene, with $D_{6+} \sim 10^{-5}$ and $D_{4+} \sim 3-6$. The compatibility of tetravalent Re, Mo and W in clinopyroxene results from the nearly ideal size of these cations ($r^{VI} = 0.66 \text{ \AA}$ for W^{4+} , 0.65 \AA for Mo^{4+} and 0.63 \AA for Re^{4+}) relative to the octahedral M1 site ($r_{0^{4+,M1}} \sim 0.66 \text{ \AA}$; Hill et al., 2000), which makes the substitution into clinopyroxene lattice energetically favorable. While the mineral/melt partitioning of Re^{4+} and Mo^{4+} were determined directly by experimental data at reducing conditions (Mallmann and O'Neill, 2007; Leitzke et al., 2017), data for W^{4+} was constrained by Fonseca et al. (2014) using a lattice strain fit to Ti, Hf, Zr and Re. This is because the extremely reducing conditions necessary to reduce all W^{6+} to W^{4+} cannot be achieved in the laboratory. The important conclusion of these studies is that the partitioning behavior of these elements in magmatic systems will be completely different depending on fO_2 , with significant implications for comparing lunar vs. terrestrial, or oceanic vs. arc magmatism.

The effect of composition on the mineral/melt partitioning of Mo, W and Re in pyroxenes has been addressed in some studies. Hill et al. (2000) and Righter and Shearer (2003) found a negative correlation between $D_{\text{Mo,W}}$ and the Ca-Tschermak (CaTs) component in clinopyroxene. However, the opposite trend was found by Leitzke et al. (2017), who observed that the clinopyroxenes with the highest CaTs reported by Hill et al (2000) were formed from melt compositions with the highest CaO. Leitzke et al. (2017) speculated that, because the activity coefficients of MoO_3 and MoO_2 in silicate melts decrease with increasing CaO content (O'Neill and Eggins, 2002), the negative correlation found by Hill et al. (2000) could be due to the high CaO content of the silicate melt rather than pyroxene composition. Leitzke et al. (2016) investigated the effect of TiO_2 in the silicate melt, observing a positive correlation between the TiO_2 content of the melt on the partitioning of Mo^{4+} and W^{4+} , but little to no effect on the partitioning of Mo^{6+} and W^{6+} (perhaps because these cations are very incompatible). Conversely, partitioning coefficients for Hf and Zr decrease with TiO_2 content of the melt. (D_{Ti} remained constant, as expected for Henry's Law behavior). Leitzke et al. (2016) argued that the introduction of $CaTi^{4+}Al_2O_6$ into clinopyroxene with increasing melt TiO_2 raises elastic strain in

811 the M1 site of clinopyroxene. The increased elastic strain (which results in a tighter parabola)
 812 leads to an increase in the partitioning values of cations whose size are close to ideal for that
 813 site (e.g. W^{4+}), and a decrease in the partitioning values of those cations whose size are either
 814 smaller or larger than the ideal for that site (e.g. Zr^{4+} and Hf^{4+}).

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817 6.2. Platinum Group Elements (Ru, Rh, Pd, Os, Ir, Pt)

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819 There have been few studies investigating the redox-dependent partitioning behavior of
 820 the Platinum Group Elements (PGE) during magmatism, with most focusing on the role of base
 821 metal sulfides and, to a lesser extent, oxides. The scarcity of partitioning data arises from the
 822 low solubility of PGE in silicate melts (typically $< a few \mu\text{g/g}$), especially under reducing
 823 conditions (O'Neill et al., 1995), and from the almost unavoidable presence of sub-micrometer
 824 metallic nuggets that plague the analyses of PGEs in synthetic silicate glasses and often lead to
 825 erroneous solubility values (Ertel et al., 2008). The solubility of a PGE in a silicate melt in
 826 equilibrium with metal can give an indication of its valence state(s), as it will dissolve according
 827 to Equation 10 (with $x = 0$ for the metal). However, despite significant effort in the last 20-30
 828 years, PGE solubility results are still controversial. This leads to ambiguous interpretations
 829 regarding the PGE speciation as a function of fO_2 . For example, solubility data for Ru are used to
 830 indicate that it either dissolves as Ru^{3+} (Borisov and Nachtweyh, 1998), or as both Ru^{3+} and
 831 Ru^{4+} (Laurenz et al., 2013) in silicate melts over planetary magmatic fO_2 s. Similarly, Os appears
 832 to dissolve as Os^{3+} under most conditions, though the possibility of Os^{4+} being stable at more
 833 oxidizing conditions cannot be ruled out (Borisov and Walker, 2000; Fortenfant et al., 2006).
 834 Data for Ir are also ambiguous, indicating speciation as either Ir^{3+} , or combinations of Ir^{1+} - Ir^{3+}
 835 or Ir^{2+} - Ir^{3+} (Brenan et al., 2005; Borisov and Palme, 1995; Fonseca et al., 2011; O'Neill, 2015).
 836 Pt appears to occur exclusively as Pt^{2+} , except perhaps at very oxidizing conditions ($FMQ > 5$)
 837 where Pt^{4+} may also be stable (Ertel et al., 1999). Of all PGEs, unambiguous evidence of redox-
 838 sensitivity over magmatic fO_2 s only exists for palladium and rhodium, both of which occur as
 839 monovalent and divalent cations under typical magmatic fO_2 s (e.g. Laurenz et al., 2010;
 840 Sommer, 2014; Fig. 2). Reports of other valence states for Pd and Rh in the literature arise from
 841 interpretation of micro(nano)nugget contaminated data (e.g. Borisov et al., 1994; Ertel et al.,
 842 1999).

843

844 Of the common magmatic minerals, spinel is the only mineral capable of accommodating
 845 significant amounts of PGEs, particularly Ru, Ir and Rh (e.g. Park et al., 2017). Wijbrans et al.,
 846 (2015) reported a marked increase in D_{Rh} around FMQ in both Fe-free and Fe-rich spinels,
 847 indicating higher compatibility of Rh^{3+} relative to Rh^{2+} . Brenan et al. (2012) observed
 848 significant variations in the partitioning of several PGEs with fO_2 for Fe- and Cr-bearing spinels.
 849 However, as noted above, the partitioning of trace elements into magnetite is controlled by the
 850 Fe^{3+} and Fe^{2+} content of the melt, and it is often difficult to deconvolve this effect, and that of
 851 drastically changing site occupancies in the spinel from that of valence state changes in the trace
 852 element of interest. Nevertheless, the data in the Fe-free system are consistent with a change
 853 from Rh^+ to Rh^{2+} , and also with olivine/melt partitioning data that shows decreasing D_{Rh} with
 854 increasing fO_2 (Brenan et al., 2003). Of the very few spinel/melt partitioning data available, Pd
 855 appears to be incompatible in spinel (Brenan et al., 2012).

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857

858 7. Concluding remarks

859

860 In this chapter we have shown that the partitioning behavior of redox-sensitive trace
 861 elements will almost inevitably change with fO_2 in magmatic systems. This is because the
 862 incorporation of cations into crystal lattices (at constant pressure, temperature and
 863 composition) depends on their charge, size, and crystal-field stabilization energies, all of which

864 likely vary substantially between the oxidized and reduced species of the same element. This
865 dependence on fO_2 means that petrologic/geochemical modelling involving elements such as Ti,
866 Cr, V, Ce, Eu, U, Mo, Re, W, Pd, Rh (and probably others) must account for the oxidation state of
867 the magma. As the only major-component redox-sensitive elements in many magmatic systems,
868 the partitioning behavior of Fe is more likely to control rather than to reflect fO_2 . Given the large
869 range (10-15 order of magnitude; Fig. 2) of magmatic fO_2 in planetary systems, significant
870 differences in the partitioning behavior of individual elements ought to be expected. For
871 instance, several studies have shown that certain trace elements behave in completely different
872 ways during lunar compared to terrestrial magmatism, or during magmatism in mid-ocean
873 ridge compared to subduction zone settings.

874

875 Nevertheless, if the partitioning of redox-sensitive trace elements is determined
876 experimentally as a function of fO_2 (and the effects of other parameters such as pressure,
877 temperature and composition accounted for), then partitioning relations preserved in natural
878 samples (e.g. glass and phenocryst in a volcanic rock) may provide useful ways of estimating the
879 oxidation state of magmatic systems. This approach has the greatest advantage of being
880 insensitive to late-stage degassing, charge-transfer (electron-exchange) reactions, or surficial
881 alteration, all of which may potentially alter other magmatic redox proxies (e.g. $Fe^{3+}/\Sigma Fe$).

882

883 The preceding discussion demonstrates the importance of a good understanding of
884 redox in modelling or interpreting trace element concentrations in melts and minerals.
885 However, it should also be apparent that further work is required to bring our ability to model
886 the partitioning of redox-variable elements. First, there are several elements for which very
887 little information is available: most notably the PGEs, which have been investigated in
888 surprisingly few minerals and whose redox transitions are poorly constrained. Copper was
889 excluded from this review because of a paucity of data, but the data of Liu et al. (2014, 2015)
890 indicate that its partitioning changes significantly from FMQ+1 to FMQ+5, a range that is highly
891 relevant to arc environments where essentially all the world's economic Cu deposits are located.
892 Likewise, experimental partitioning data by Mallmann and O'Neill (2009) indicate a potential
893 change of phosphorus valence states from 5+ to 3+ under conditions below FMQ-4, but the
894 results could also be explained by volatility loss from the melt after crystallization. And, under
895 ultra-reducing conditions (\sim FMQ-7), where Nb and Ta begin to show siderophile tendencies,
896 there is evidence that they may take on lower valence states in silicate melts (Cartier et al.,
897 2014).

898

899 Furthermore, the dependence of these redox transitions and partition coefficients on
900 temperature and melt composition is complex, and only partially understood for the majority of
901 the redox couples addressed here. As seen in Figure 4, melt composition affects the partitioning
902 of different oxidation states of the same element in differing ways, an effect that is additional to
903 the influence of stoichiometric controls on trace element partitioning. Similarly, melt
904 compositional effects can shift the redox transition by up to 6 log units of fO_2 (Fig. 2). Because of
905 these complications, extreme caution should be taken when attempting to study partitioning of
906 redox-variable elements outside their calibrated ranges.

907

908 Many factors are important in selecting a mineral/melt partitioning oxybarometer. First,
909 the anticipated redox state of the system will give an indication of which redox couples are
910 relevant, but a large contrast between the compatibilities of the valence states of the element
911 can sufficiently outweigh this consideration (for example, zircon will develop a Ce anomaly at
912 conditions that are many orders of magnitude more reducing than the midpoint of the Ce^{3+} - Ce^{4+}
913 equilibrium). The figures presented in this chapter may guide the geoscientist in determining
914 whether the partition coefficient is likely to be sensitive to fO_2 for her or his samples. As
915 emphasized above, the availability of appropriate experimental constraints is important, as are
916 more general petrological considerations, for instance resistance to alteration processes such as

917 diffusive re-equilibration. In the ideal case, multiple elements and minerals would be combined
918 to check for internal consistency.

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920

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1353 Figure captions

1354

1355 **Figure 1. (a)** Schematic illustration of lattice strain parameters D_0 , r_0 and E for the mineral/melt
1356 partition coefficient (D) of an element i as modelled by the Brice equation (Blundy and Wood,
1357 1994); $D_i = D_0 \exp([-4\pi EN_A(r_0 - r_i)^2 - 1/3(r_0 - r_i)^3]/RT)$, and **(b, c, d)** measured changes in
1358 these parameters with cation valence state (i.e. charge) as determined experimentally for
1359 CaSiO_3 /melt partitioning (Law et al., 2000).

1360

1361 **Figure 2.** Relative oxygen fugacity conditions (in log units relative to FMQ, the fayalite-
1362 magnetite-quartz buffer; values from O'Neill, 1987) for several planetary geochemical
1363 reservoirs (Frost and McCammon, 2008; Grossman et al. 2008; Wadhwa, 2008), and valence
1364 state transition ranges for the elements discussed in this chapter. See text for references. Also
1365 plotted for reference the IW (Fe-FeO), NNO (Ni-NiO), and HM (hematite-magnetite) buffers.

1366

1367 **Figure 3:** Mineral/melt partitioning of transition metals as a function of oxygen fugacity (in log
1368 units relative to FMQ, the fayalite-magnetite-quartz buffer; values from O'Neill, 1987) for
1369 selected minerals. Abbreviations: amph - amphibole, cpx - clinopyroxene, opx - orthopyroxene,
1370 olv - olivine, plag - plagioclase feldspar, spl - spinel, zrc - zircon. All curves shown in the main
1371 diagram are best fits of the experimental data to an equation with the same form as Equation
1372 22. The fraction (from 0 to 1) of the different valence states of each element are plotted on top
1373 of the diagram. These were calculated using Equation 13 using available data. **(a)** Fe
1374 mineral/melt partitioning data from King et al., 2000 (amph), Mallmann and O'Neill, 2009 (spl,

1375 olv, opx, cpx) and Phinney, 1992 (plag). **(b)** Cr mineral/melt partitioning data from Mallmann
 1376 and O'Neill, 2009 (olv, opx, cpx). **(c)** Ti mineral/melt partitioning data from Mallmann and
 1377 O'Neill, 2009 (olv, opx, cpx), Peters et al. 1995 (plag) and Burnham and Berry, 2012 (zrc). **(d)**
 1378 **(d)** V mineral/melt partitioning data from Mallmann and O'Neill, 2009 (olv, opx, cpx, spl).
 1379

1380 **Figure 4.** Comparison of Cr olivine(Fo_{100})/melt partitioning coefficients obtained at 1 bar and
 1381 1320 °C as a function of oxygen fugacity (in log units relative to FMQ, the fayalite-magnetite-
 1382 quartz buffer; values from O'Neill, 1987) for three Fe-free compositions in the CaO-MgO- Al_2O_3 -
 1383 SiO_2 system: FAS1, FAD1 and FAD3 (Hanson and Jones, 1998). The results illustrate a mild effect
 1384 of melt composition on the partitioning of Cr^{3+} yet no significant effect on the partitioning of
 1385 Cr^{2+} . The fraction (from 0 to 1) of the different valence states are plotted on top of the diagram.
 1386

1387 **Figure 5.** Zircon/melt partition coefficients for REE elements determined experimentally at 1
 1388 bar as a function of oxygen fugacity by Burnham and Berry (2012). The colored segments
 1389 highlight systematic changes in the partitioning of Ce and Eu with oxygen fugacity (labelled
 1390 relative to the FMQ buffer).
 1391

1392 **Figure 6.** Mineral/melt partitioning of redox-sensitive rare earth elements (REE), Ce **(a)** and Eu
 1393 **(b)** as a function of oxygen fugacity (in log units relative to FMQ, the fayalite-magnetite-quartz
 1394 buffer; values from O'Neill, 1987) for selected minerals. Abbreviations: bdd - baddeleyite, cpx -
 1395 clinopyroxene (aluminous diopside), olv - olivine, opx - orthopyroxene, plag - plagioclase
 1396 feldspar, ttn - titanite, zrc - zircon. Where data were not obtained as a function of $f\text{O}_2$, partition
 1397 coefficients for Ce^{3+} , Ce^{4+} , Eu^{2+} and Eu^{3+} (or estimates thereof) were used in conjunction with
 1398 the $\text{Ce}^{4+}/\Sigma\text{Ce}$ and $\text{Eu}^{3+}/\Sigma\text{Eu}$ models of Burnham and Berry (2014) and Burnham et al. (2015).
 1399 Data from Klemme and Meyer, 2003 (bdd), Leitzke et al., 2017 (cpx-Ce, plag-Ce, olv), Sun et al.,
 1400 1974 (cpx-Eu, plag-Eu, opx-Eu), van Kan Parker et al., 2011 (opx-Ce), Bachmann et al., 2005
 1401 (ttn), and Burnham and Berry, 2012 (zrc). See Fig. 3 for more details.
 1402

1403 **Figure 7.** Mineral/melt partitioning of U as a function of oxygen fugacity (in log units relative to
 1404 FMQ, the fayalite-magnetite-quartz buffer; values from O'Neill, 1987) for selected minerals.
 1405 Abbreviations: cpx - clinopyroxene (diopside), olv - olivine, opx - orthopyroxene, zrc - zircon.
 1406 Data from Fonseca et al., 2014 (opx, cpx, olv), and Burnham and Berry, 2012 (zrc). See Fig. 3 for
 1407 more details.
 1408

1409 **Figure 8.** Mineral/melt partitioning of siderophile elements (Mo, W and Re) as a function of
 1410 oxygen fugacity (in log units relative to FMQ, the fayalite-magnetite-quartz buffer; values from
 1411 O'Neill, 1987) for selected minerals. Abbreviations: cpx - clinopyroxene, opx - orthopyroxene,
 1412 olv - olivine, spl - spinel, grt - garnet. Data from Leitzke et al., 2017 (Mo: cpx, opx, olv), Wijbrans
 1413 et al., 2015 (Mo: spl), Fonseca et al., 2014 (W: opx, cpx, olv), and Mallmann and O'Neill, 2007
 1414 (Re: cpx, opx, spl, grt, ol). See Fig. 3 for more details.

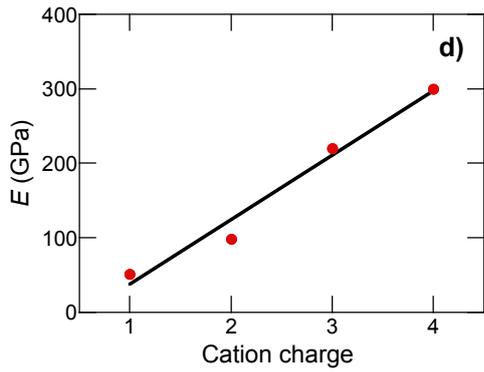
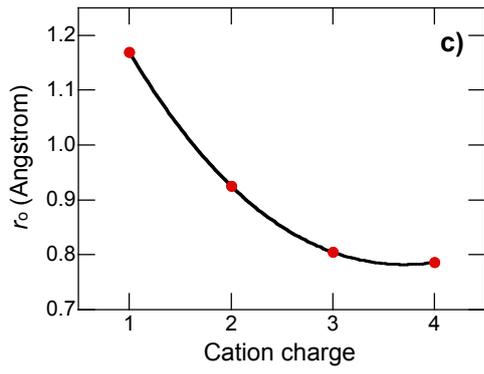
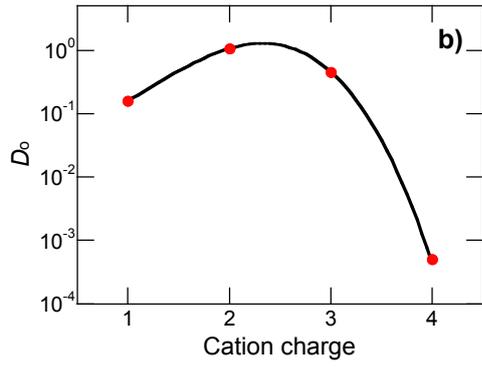
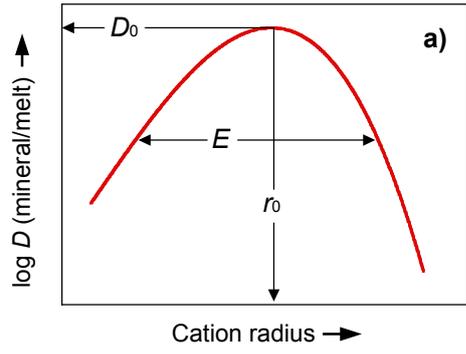


Figure 2

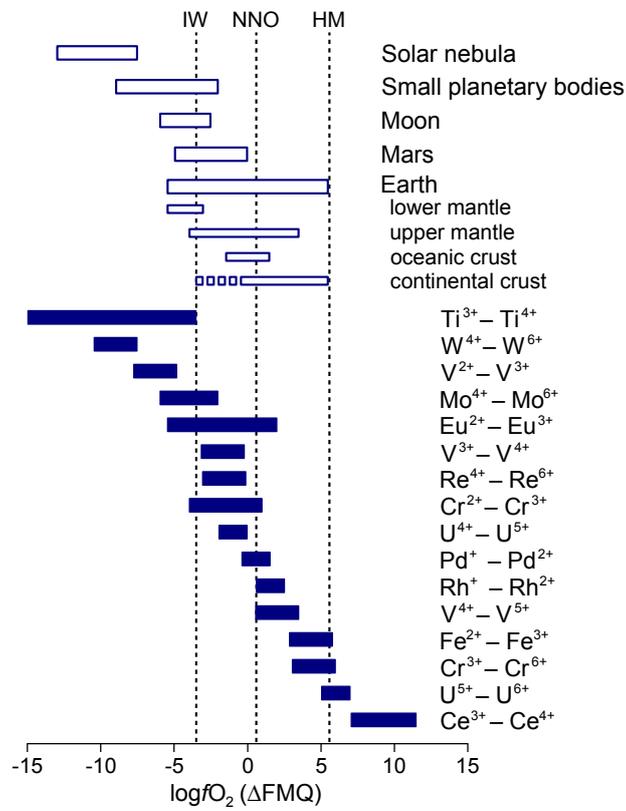


Figure 3

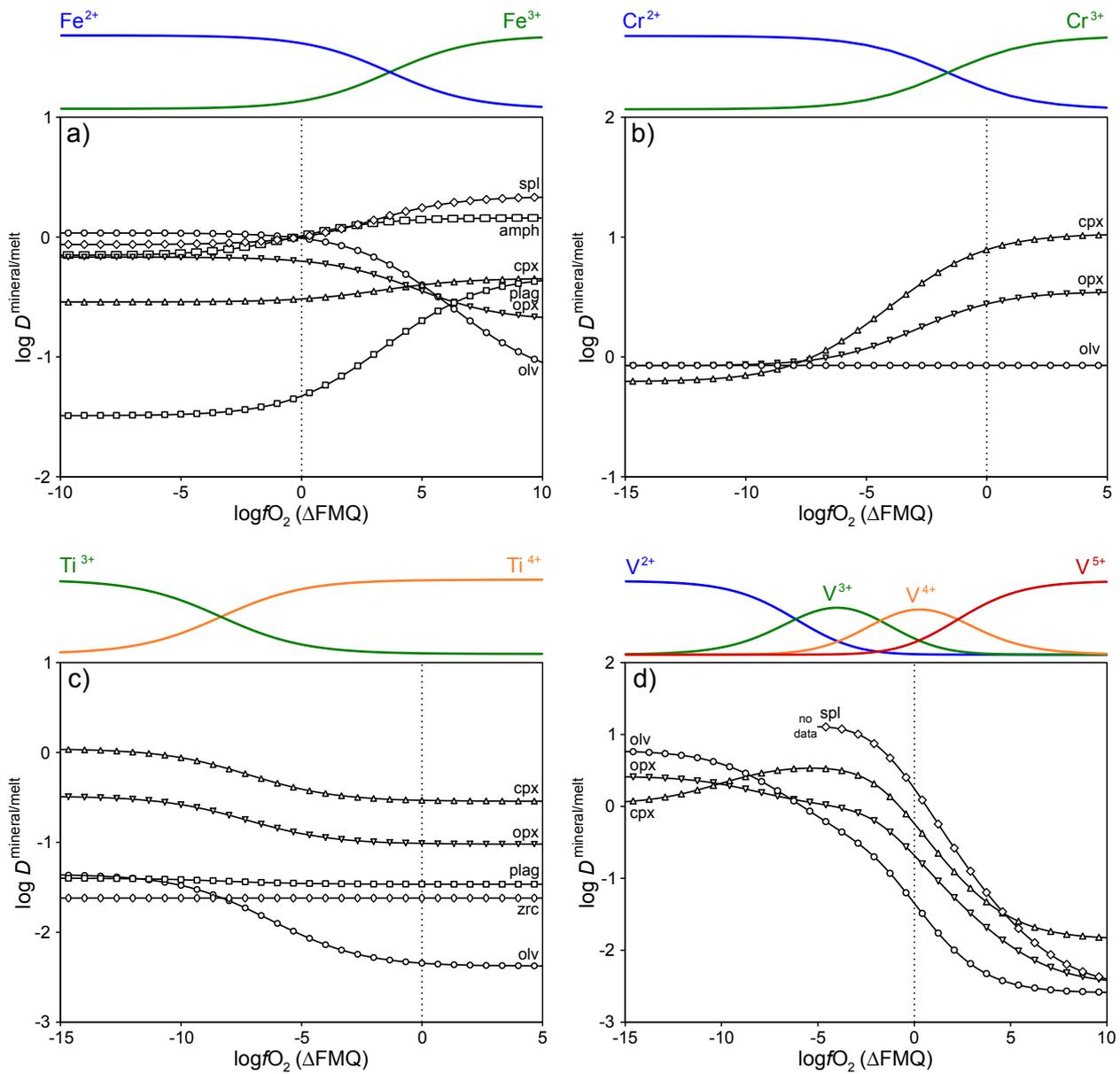


Figure 4

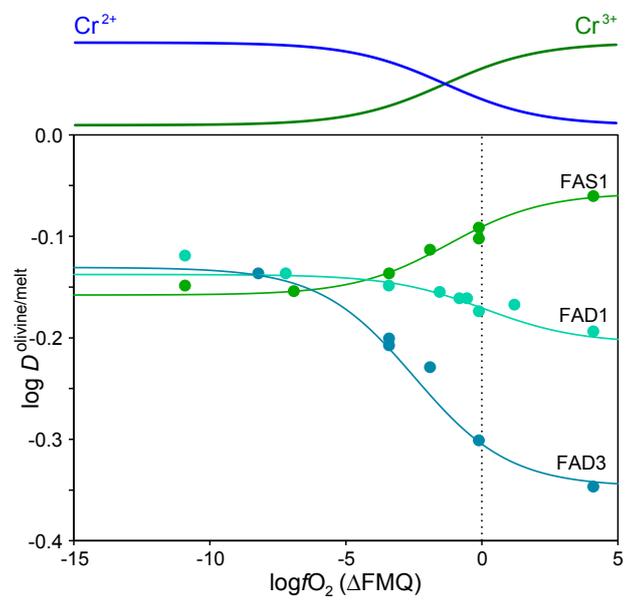


Figure 5

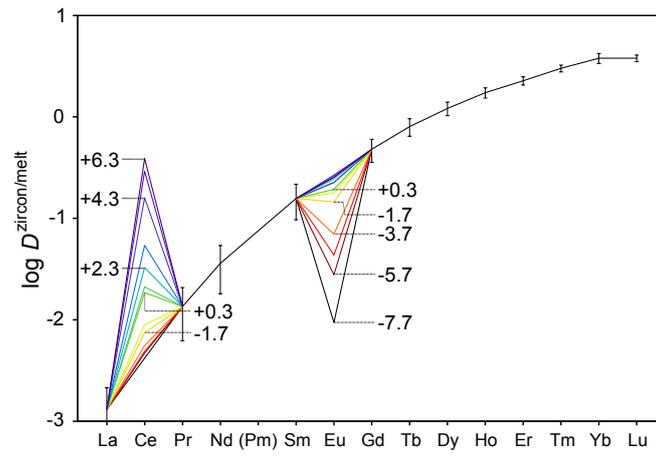


Figure 6

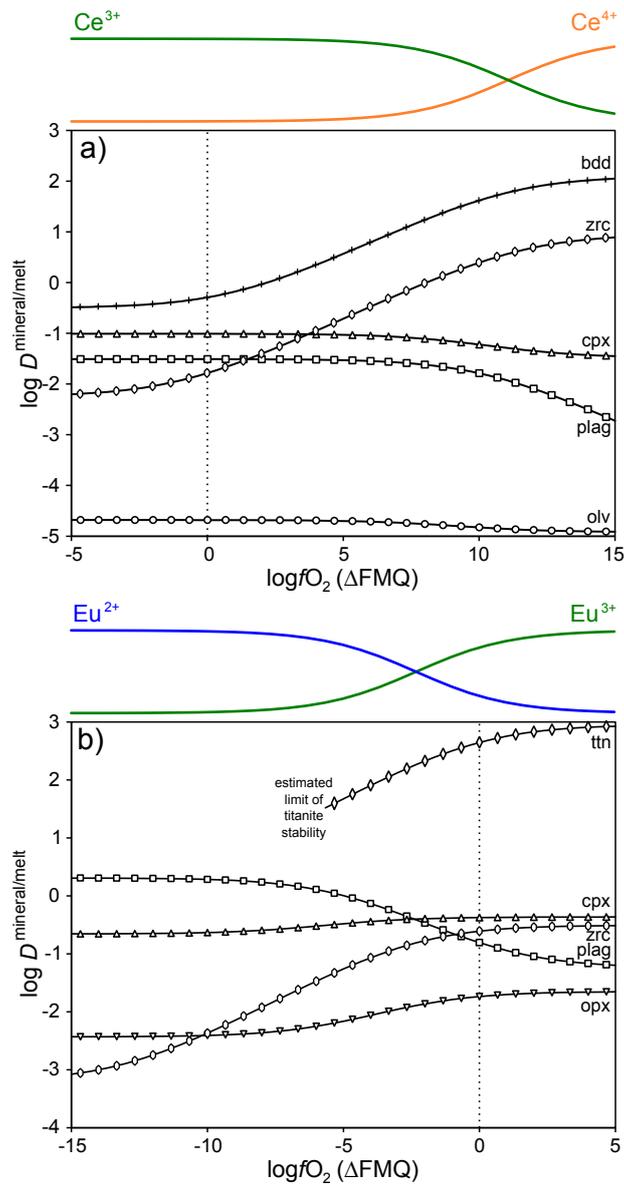


Figure 7

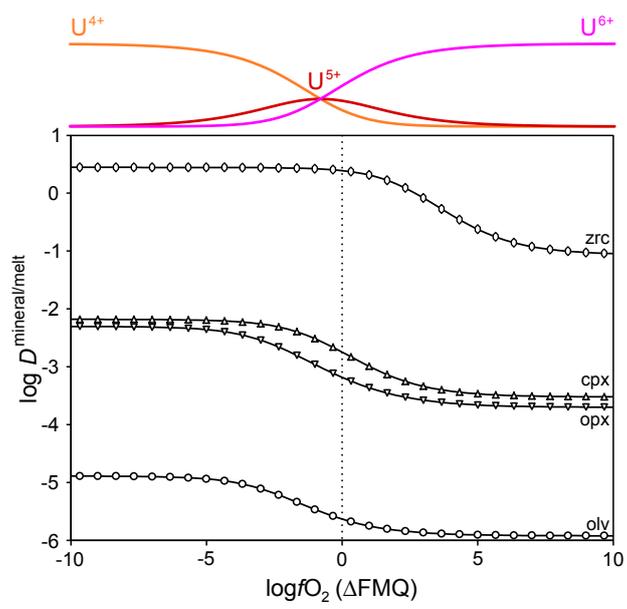


Figure 8

